

# HYPORHEIC FLOW IN A MOUNTAINOUS RIVERINE SYSTEM

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# ABSTRACT

Investigation into the effects of beaver dams on hyporheic exchange in peat dominated mountainous streams is needed to better understand stream-floodplain connections and improve our overall conceptual model of water storage and flow through riverine valleys. The objective of this study was to determine the influence of in-stream beaver dams on vertical and lateral hyporheic exchanges. Hyporheic interactions were examined using hydrometric methods to determine both flow pathways and water fluxes for a second-order stream draining a Canadian Rocky Mountain peatland. Investigation was conducted on two instream beaver dams and an undammed reference section for the ice free periods of summers 2006 and 2007 at the Sibbald Reseach Basin located in Kananaskis Country, Alberta, Canada. Lateral hyporheic fluxes dominated over vertical hyporheic fluxes, due to a layer with low saturated hydraulic conductivity ( $K \sim 10^{-7} - 10^{-9}$  m/s) just below the streambed throughout most of the study reach. The lateral flow around the north dam ( $> 0.6$  m high) resulted in fluxes that ranged from 0.002 to 0.015 L/s in the near bank environment. These results confirm that hydraulic properties of the substrata are an important factor in the development of hyporheic exchange in stream systems draining peatlands. Results also demonstrate the ability of beaver to connect valley floors to their streams, which maintains seasonally stable water tables and wetland conditions in the riparian zone.

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To Daryl and Amalia. Thank you for waiting.

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# LIST OF ABBREVIATIONS

$\alpha$	Dimensionless empirical parameter
$C_b$	Background concentration of tracer (ppm)
$C_g$	Tracer concentration of groundwater (ppm)
$C_i$	Concentration of injected tracer (ppm)
$C_p$	Plateau concentration (ppm)
$C_s$	Tracer concentration in the stream (ppm)
$C_w$	Tracer concentration in a well (ppm)
FPC	Flood Pulse Concept
$\Delta h$	Change in head (m)
HCC	Hyporheic Corridor Concept
$K$	Hydraulic conductivity (m/s)
$\Delta l$	Length over which there is a change in head (m)
$L$	Length (m)
$L_e$	Length of well screen (cm)
MODFLOW	Two- or three-dimensional groundwater model
MODPATH	Particle tracking package
MT3D99	Transport simulation model
$n$	Number of stream tubes
OTIS-P	One-dimensional surface water solute transport model
$Q$	Flux (L/s)
$q$	Specific discharge (m/s)
$Q_b$	discharge at the base of the study reach (ppm/min)
$\Delta q_{gw}$	Discharge per unit length (m <sup>2</sup> /s)
$q_{gw}$	Total discharge per unit stream length (m <sup>2</sup> /s)
$Q_h$	Discharge at the head of the study reach (ppm/min)
$Q_i$	Flow rate of injectate (ppm/min)
$Q_0$	Discharge from tracer test (ppm/min)
$r_c$	Well radius (cm)
$R_e$	Effective radius of slug test (cm)
$r_s$	Radius of well screen (cm)
$S$	Standard deviation
SDC	Serial Discontinuity Concept
$SE$	Standard error
$S_s$	Specific storage (cm <sup>-1</sup> )
SU	Stick-up
$SW$	Percent surface water (%)
$T_m$	Minimum travel time (min)
$T_0$	Time to a normalized head of .368 (s)
VHG	Vertical hydraulic gradient
ZONEBUDGET	Hyporheic flux model

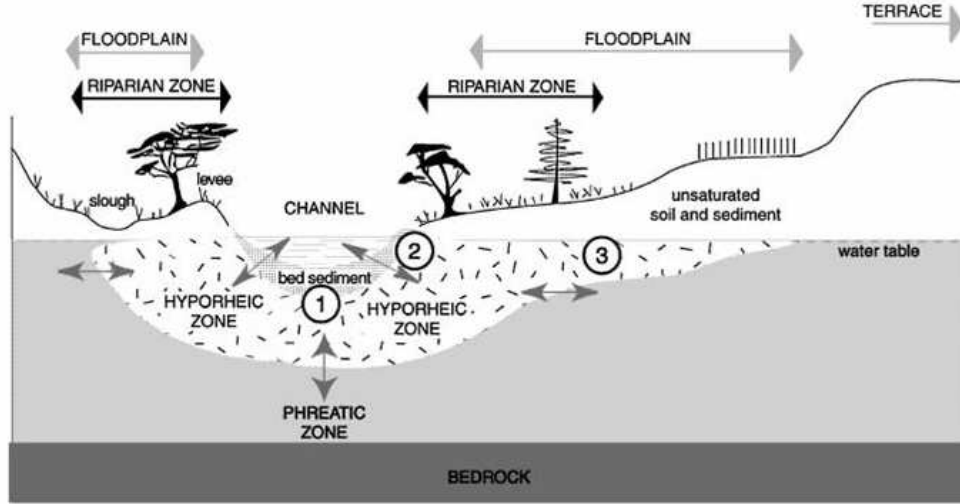
# CHAPTER 1

## INTRODUCTION

Sustainable management of water resources requires conceptual and physical knowledge of hydrological processes. Hyporheic exchange is potentially one of the important hydrological processes that should be considered in water management schemes. It is also a key process to consider in the development of conceptual riverine frameworks because it is driven by physical and biological processes that exist ubiquitously in aquatic systems. Hyporheic exchange takes place in a dynamic zone between the stream and groundwater system known as the hyporheic zone. Within the hyporheic zone water exchange is vertical and lateral, having hydrological, chemical, and biological importance. Research on hyporheic systems is currently taking place in hydrology and ecology and the physical and conceptual knowledge base is expanding. Little of the recent scientific work has focused on incorporating impacts of biological processes on flow within the hyporheic zone and thus, is the focus of this thesis.

### 1.1 Defining Hyporheic Processes

One of the most basic processes taking place at the stream-groundwater interface is the recharge and discharge of stream water to the local aquifer. Without this recharge and discharge interaction, surface and groundwater would either flow quickly out of a drainage basin via surface streams or seep slowly into the near bank zone and assimilate into the deeper regional groundwater system (Winter, 1999; Konrad, 2006). Research on hyporheic flow has been conducted in both hydrology and ecology and both groups have produced functional and theoretical definitions for the surface-



**Figure 1.1:** A conceptual diagram outlining the three zones of hyporheic exchange: (1) vertical hyporheic exchange, (2) horizontal hyporheic exchange, and (3) the floodplain hyporheic exchange zone (Steiger et al., 2005).

groundwater exchange, as well as defined spatially discrete zones or regions where this exchange takes place.

The term hyporheic is used to represent both the interface and the process where surface water and groundwater are exchanged through the channel bank and bed in the near river-stream environment. Lautz and Siegel (2006) defined the hyporheic zone as the “interface between surface and groundwater in near-stream sediments”. This definition is simple and clear; hyporheic fluxes are defined to take place at any point where the surface and groundwater are adjacent to one another in the near-stream environment. This definition does not, however, give any indication whether or not this zone is a static, distinct area, or if it is a dynamically variable zone.

These shortcomings are overcome in the definition given by Boulton et al. (1998), who define the hyporheic zone as a “spatially fluctuating ecotone between the surface stream and the deep groundwater”. Boulton et al. (1998) even go on to characterize some of the processes and factors which take place in this active zone, such as “important ecological processes and their requirements and products[,] [which] are influenced at a number of scales by water movement, permeability, substrate particle

size, resident biota, and the physiochemical features of the overlying stream and adjacent aquifers.” This definition seems to encompass all the possible factors that are needed to define the hyporheic zone, but the definition states that the interaction is with the surface stream water and deep groundwater. This may be a bit restrictive if hyporheic zones and interactions are present in streams located within wetlands where groundwater tables are shallow or right at the surface. Thus, this definition is also unsatisfactory.

Perhaps the most functional definition is that of Cardenas et al. (2004). They define the hyporheic zone as “an area where water infiltrates from streams then flows through streambed sediments and stream banks and returns to the surface after relatively short pathways.” This definition adds to the clarity of hyporheic exchange by adding a relative temporal framework.

From these definitions a plausible idea of where hyporheic exchange takes place, and what its processes are, can be made clear. In this thesis, hyporheic exchange is therefore defined as the mixing and transient transport of surface water and groundwater; the hyporheic zone spatially fluctuates and is in the near-stream bank/bed sediment environment (see Figure 1.1. This study will not, however, address zone 3).

# CHAPTER 2

## REVIEW OF THE LITERATURE

This chapter will give a general review of the literature pertaining to hyporheic exchange, establishing the background information needed to assess the gaps within this field of research. The background knowledge will then be used as a springboard from which the thesis objectives were formulated.

### 2.1 Introduction

Groundwater and streams are both important water resources. Over one-sixth of the Earth's population relies on these two resources for potable water supplies (Barnett et al., 2005). Thus, ensuring the availability and quality of these resources is of the utmost importance. Management and responsible use are keys to the sustainability of groundwater and surface water assets. Proper and sustainable resource management requires a conceptual and physical understanding of the processes regulating the availability of surface and groundwater systems. Understanding the conceptual and physical components of a process requires investigation into the mechanical, chemical, and thermal interactions within the process' geological and ecological settings.

There has been an increase in the number of studies researching water flow across the interface between surface and groundwater systems due to its importance to the hydrological functioning, chemistry, temperature, and solute transport in both groundwater and surface water systems (Boulton et al., 1998). The interface between groundwater aquifers and surface water systems has been coined the hyporheic zone and the process is referred to as hyporheic exchange. The research conducted at any interface has the potential to increase the conceptual understanding of the processes

within it in ways that studying a system in isolation cannot. Such studies then increase the ability to predict or understand the overall system wherein the intersection lies. The interface between stream surface waters and nearby shallow aquifers can be characterized by fluvial recharge of aquifers or by groundwater seepage back into the surface stream system, all within the near-stream environment (Konrad, 2006).

The remainder of this section will establish a theoretical framework for hyporheic exchange and the standard ways in which to investigate this process, explore the major contributions of contemporary research, and discuss gaps in the theory or literature, as well as outline potential directions for future research.

## 2.2 Theoretical/Conceptual Framework

Hyporheic exchange is a medium-to-small (tens of meters to centimeters) scale process, which takes place within the overall hydrological cycle of watersheds. In drainage basins, precipitation in the form of either snow or rain infiltrates the hillslopes and percolates into the ground, where it flows along one of two basic flow paths that are regulated by hydrodynamic controls (Triska et al., 1993b). Shallow groundwater flow paths exist under saturated antecedent soil moisture conditions, or in areas where shallow impermeable strata layers impede the downward percolation of water (Hood et al., 2006). In this situation, groundwater systems can transport water to the channel at the maximum rate of the porous media's saturated hydraulic conductivity. Longer groundwater flow paths also transport hillslope water towards the channel when water is able to percolate down to the water table, but the transport time is longer.

Once groundwater and surface water reach the floodplain, these water sources become part of a dynamic, complex system of groundwater and stream water flow patterns. The complexity of the floodplain hydrology is partly controlled by the processes of hyporheic exchange, where groundwater and surface water are transported and mixed, creating unique chemical and hydrological environments (Boulton et al., 1998). Hyporheic exchange complicates the simplistic view of a drainage basin func-

tioning as a topographic system, where groundwater from the hillslopes flows to the slope base into the floodplain and toward the channel. Instead, hyporheic flow induces both lateral and vertical exchange between the stream and groundwater systems.

The theoretical framework of the hyporheic process is situated in both hydrology and stream ecology fields. The hydrological process of hyporheic exchange fits into four broad ecological and hydrological concepts: bank storage, flood pulse, hyporheic corridor, and serial discontinuity. Each of these will be discussed below.

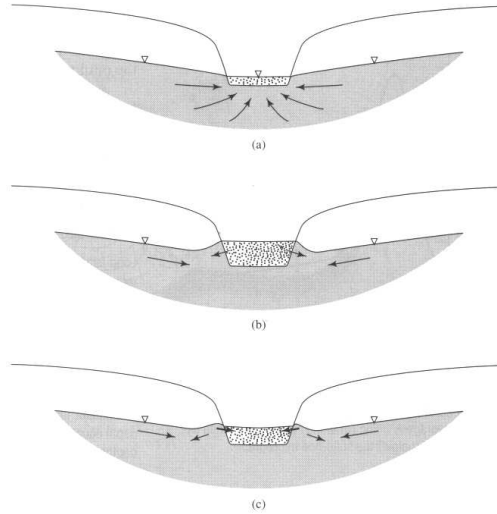
### **2.2.1 Bank Storage**

The lateral exchange of stream and groundwater is often referred to as bank storage and is mainly studied by hydrologists. The concept of bank storage involves the exchange and storage of flood waters in the stream banks and floodplain during high flow events (Whiting and Pomeranets (1997); Figure 2.1b). This stored water is then slowly released back into the stream as the stream stage drops below the subsurface water table (Whiting and Pomeranets (1997); Figure 2.1a). Thus, the concept of bank storage allows for the stream bank and floodplain to attenuate flood waters during large magnitude events (Whiting and Pomeranets, 1997). As a result, understanding the function and importance of hyporheic processes will contribute to the conceptual and theoretical knowledge of bank storage. Mostly, this process is studied through numerical models rather than directly studied *in situ*.

### **2.2.2 Flood Pulse**

Hyporheic processes can be found in the larger conceptual theory of the Flood Pulse concept (FPC); a theory created by stream ecologists. The FPC describes a lateral, surface connectivity between the floodplain and surface stream systems, where over-bank flows allow for nutrient deposition and aquifer recharge on the floodplain (Junk et al., 1989). This concept moves away from the simple idea of longitudinal flow of water down the channel, but instead implies that there is a flux of stream water to





**Figure 2.1:** A conceptual diagram illustrating the phases of bank storage: (a) stream recharge from base flow, (b) groundwater recharge from an increase in stream stage, and (c) stream recharge from a local groundwater mound (Dingman, 2002).

the near-stream environment. Hyporheic flow is the medium scale process of the FPC where groundwater does not just simply flow from the hillslopes to the channel and then preferentially down stream, but instead has a reverse lateral connection to the floodplain subsurface.

### 2.2.3 Hyporheic Corridor Concept

The Hyporheic Corridor Concept (HCC) discussed in Stanford and Ward (1993) extends the concept of the flood pulse to include a vertical component to the theory. The drivers in HCC are geomorphology, head gradients, and solute gradients. These drivers induce vertical exchange of surface and subsurface water through short flow paths in the bed. In this concept, the flux of groundwater is an important source for regulating summer stream temperatures. As well, the exchange of oxic surface water to the anoxic groundwater environments creates pockets of increased microbial activity in the hyporheic zone. The HCC develops the hydrogeological importance of hyporheic fluxes where groundwater flow patterns are affected and geochemical reactions are altered.

### **2.2.4 Serial Discontinuity**

Hyporheic exchange also fits into the Serial Discontinuity Concept (SDC) (Ward and Stanford, 1995). In the SDC, streams do not change uniformly in the longitudinal directions of drainage basins as described in the river continuity concept. For example, it is accepted that headwaters are confined reaches due to their geological setting and that lower stream reaches are basically unconfined and very broad. The SDC states that there is no clear progression from the confined headwater to the broad, low lying reaches down the fluvial network. Instead, hydrological and ecological patterns are variable and patchy, making the floodplain width non-uniform down the stream network. Hyporheic processes are empirical evidence for this concept. Areas of high and low hyporheic exchange do not vary in a uniform manner down the length of a channel; instead, hyporheic processes are variable at all scales and are primarily controlled by basin geology, streambed and bank morphology, and biological processes (Kasahara and Wondzell, 2003; Cardenas et al., 2004; Lautz et al., 2006).

## **2.3 Methods for Measuring Hyporheic Exchange**

In order to assess the importance and effects of hyporheic exchange on the hydrology of groundwater systems, process-based field studies are needed. This research then needs to be incorporated into predictive hydrological models. There are four main methods used to look at flow patterns in the hyporheic zone: hydrometric instrumentation, tracer tests, detailed discharge measurements and water temperature methods; all of which can then be used to model hyporheic interactions.

### **2.3.1 Hydrometric**

Hydrometric methods used in assessing hyporheic flow implement groundwater monitoring wells and in-stream and bank piezometers in order to directly observe changes in groundwater storage and flow (Woo and Waddington, 1990). The changes in flow and groundwater storage lead to the identification of areas within the stream reach

where groundwater discharge and recharge are taking place. These areas identify the locations where active hyporheic exchange may be changing the chemistry and temperature of the stream in that location, as well as redirecting nearby groundwater flow patterns.

When employing hydrometric methods, it is necessary to obtain hydraulic conductivity values at each well or piezometer. Hydraulic conductivity values can be estimated by bail or slug type tests, which provide a point approximation for conductivity (Freeze and Cherry, 1979). Once spatially distributed values of conductivity are found, and if an evenly distributed amount of information about the average water table elevation is available, one could implement the use of flow nets to represent the major groundwater flow features in the hyporheic zone. Hydraulic conductivity and hydrometric measurements can be used to calculate vertical and horizontal hydraulic gradients (Cey et al., 1998; Baxter et al., 2003). Vertical hydraulic gradients (VHG) can be determined by,

$$VHG = \frac{\Delta h}{\Delta l}, \quad (2.1)$$

where  $\Delta h$  (m) is the difference in head between the water level in a piezometer and the level of the stream surface and  $\Delta l$  (m) is the distance from the streambed surface to the first opening in the piezometer. Vertical discharge ( $Q$  (m<sup>2</sup>/s)) is then found using a modified Darcy's law,

$$Q = K \frac{\Delta h}{\Delta l} w, \quad (2.2)$$

for which  $K$  (m/s) is the vertical hydraulic conductivity and is denoted as a positive value so that discharges from the stream are defined to be negative and discharges into the stream are then positive, and  $w$  (m) is the width of the stream. Using Darcy's law where the infinitesimal head loss at a certain point,  $dh$  is due to various factors and may be written

$$dh = \frac{dp}{\gamma} + dz + \frac{d(v^2)}{2g} + \sigma d\pi_h, \quad (2.3)$$

where  $p$  (N/m<sup>2</sup>) is pressure,  $\gamma$  (N/m<sup>3</sup>) is the specific weight of water,  $v$  (m/s) is the velocity of flow,  $g$  (9.81 m/s<sup>2</sup>) is the acceleration due to gravity,  $z$  (m) is elevation,  $\sigma$  (0 for no membrane effects and 1 for a perfect membrane) is the membrane efficiency

or reflection coefficient and  $\pi_h$  (m) is osmotic head. Because velocities in porous media are usually low, velocity heads may be neglected as well as the osmotic head, which are predominantly found only in clay and shale porous media (Todd and Mays, 2005; Ingebritsen et al., 2006).

In order to assess the groundwater discharge into a reach of interest, it is convenient to first determine the discharge per unit length into the stream. Dingman (2002) gives a numerical solution to the steady-state two-dimensional form of the Laplace Equation, where the discharge per unit length of the stream  $\Delta q_{GW}$  into a stream for each streamtube can be calculated as,

$$\Delta q_{GW} = K \cdot \Delta h, \quad (2.4)$$

where a streamtube is the area between two adjacent flowlines,  $K$  (m/s) is the hydraulic conductivity, and  $\Delta h$  (m) is the head difference between the stream and the closest up-gradient equipotential line. The total discharge per unit stream length ( $q_{GW}$  (m<sup>2</sup>/s)) for a reach is given by,

$$q_{GW} = n \cdot \Delta q_{GW}, \quad (2.5)$$

where  $n$  is the number of streamtubes discharging to the stream. From here, the total flow ( $Q$  (m<sup>3</sup>/s)) into a stream for a reach can be found by,

$$Q = L \cdot q_{GW}, \quad (2.6)$$

where  $L$  (m) is the length of the reach. Negative values provided by this method indicates flow from the stream into the hyporheic zone and are useful to demonstrate the type of information that can be gained from this methodology. This method is however, highly dependant on hydraulic conductivity, for which it can be difficult to obtain accurate values due to its variability in highly heterogeneous landscapes (Cey et al., 1998).

### 2.3.2 Temperature

The temperature difference between surface water and groundwater systems can be used as a type of tracer to estimate hyporheic exchange (Wondzell and Swanson,

1996b; Dingman, 2002). This temperature difference is partly due to the surface water's exposure to ambient air temperatures. The temperature gradients between the two systems can be used to calculate the mixing, relative proportion, or the direction of surface-groundwater flow in the near-streambed sediments of a channel (Wondzell and Swanson, 1996b; Dingman, 2002). The temperature method can be combined with a simple mixing model in order to calculate the percentage of stream and groundwater within wells (Wondzell and Swanson, 1996b). However, temperature is not a conservative tracer; it is affected by heat stored in the sediments and is highly variable on daily and seasonal time scales (Wondzell and Swanson, 1996b; Dingman, 2002). Although this method is useful in determining the vertical fluxes of water within the stream channel, it is not easily transferred to the bank environment for use in assessing the lateral exchange and spatial extent of hyporheic flow patterns (White, 1990). This is due to the errors that can be associated with taking groundwater temperature measurements in wells, such as the exposure of groundwater to the ambient air temperature, and influences of the conduction of heat in the well itself.

### **2.3.3 Tracer Tests**

Tracer tests can be used to measure hyporheic storage parameters, flow paths, and fluxes (Lautz and Siegel, 2006). The migration of tracers are monitored from observation locations downstream of the injection point (Lautz et al., 2006). Tracer tests can be conducted with any chemical tracer that is “(1) present in measurable amounts ... [and is] highly soluble; (2) not taken up or released in the vadose zone; and (3) not taken up or released by vegetation” (Dingman, 2002). Common chemical tracers that are used in hydrological studies include chloride and bromide (Triska et al., 1993a; Dingman, 2002). Artificial tracers such as rhodamine can be injected into a flow network, or naturally present tracers such as chloride can be implemented. Neither of these are commonly used by plants, nor are they involved in soil-chemical reactions, which makes them acceptable tracers (Triska et al., 1993a; Dingman, 2002). The difference between natural and artificial tracers is that a

natural tracer is present in both the surface and groundwater system at some concentration, but the concentrations are sufficiently different that each source can be characterized by that tracer. Artificial tracers are not significantly present in either system and are injected into one until a known concentration is reached and the rise in concentration in the uncontaminated source indicates the amount of mixing.

Determining hyporheic flow patterns using tracer methods has been described by Triska et al. (1993a,b). The percentage of surface water present in the hyporheic zone and the discharge of groundwater can be quantified by observing the transport of a tracer injected into a stream reach from the head to the base. The time for the tracer concentration to rise above background levels at the base is considered the minimum travel time,  $T_m$  (min). The time that it takes the tracer to reach a plateau concentration  $C_p$  (ppm) at the base is  $T_p$  (min). Once the concentration reaches the plateau, discharge at the head  $Q_h$  (ppm/min) and base  $Q_b$  (ppm/min) can be calculated using their plateau concentration values. The difference between  $Q_h$  and  $Q_b$  is the groundwater input to the stream. The equation developed by Triska et al. (1993b) for discharge  $Q_0$  (ppm/min) is,

$$Q_0 = \frac{C_i - C_b}{C_p - C_b} Q_i, \quad (2.7)$$

where  $C_i$  (ppm) is the concentration of injected tracer,  $C_b$  (ppm) is the background concentration of tracer at site, and  $Q_i$  (ppm/min) is the flow rate of injectate. The percent surface water (SW (%)) was formulated as

$$\%SW = \frac{C_w - C_g}{C_s - C_b} 100, \quad (2.8)$$

where  $C_w$  (ppm) is the tracer concentration in the well,  $C_g$  (ppm) is the tracer concentration in groundwater, and  $C_s$  (ppm) is the tracer concentration in the stream (Triska et al., 1993b).

The lateral extent and timing of hyporheic exchange can also be monitored using tracers by comparing the nominal travel time of the tracer between well and adjacent channel location (Triska et al., 1993a). The nominal travel time of a tracer is defined as either the time it takes to attain 50% of the background-corrected plateau

concentration on the rise, or as the time it takes for the same amount of decrease on the fall. The utility of tracer tests are evident and they are readily reported in the literature (Triska et al., 1993a,b; Lautz and Siegel, 2006; Lautz et al., 2006). However, the temporal resolution of this method is restricted usually to one or two days of a field season due to the labor intensity needed to conduct such a test. Thus, this method is not useful in capturing seasonal variations in hyporheic exchange. The use of tracer methods which employ artificial tracers are also limited in their use within the boundaries of parks and other protected areas, where they are restricted.

### **2.3.4 Discharge**

Another method that can be used to explore hyporheic exchange makes use of detailed stream discharge measurements. The discharge method is implemented by taking detailed stream discharge measurements using the area-velocity technique at approximately 30 m intervals (Cey et al., 1998). These measurements allow researchers to assess the changes in discharge between an upstream and downstream segment of a channel. The measurements identify areas of increased (stream gain) or decreased (stream loss) discharge between sample points, thus allowing researchers to locate areas of hyporheic exchange. This method, however, is also quite labor intensive and therefore temporally restrictive. As well, this method magnifies errors from the area-velocity technique, which makes it unreliable at spatial scales less than approximately 30 m (Cey et al., 1998).

### **2.3.5 Models**

There are a few common models used to predict and examine the hyporheic process. OTIS-P is a one-dimensional, surface water, solute transport model that can be used to model breakthrough curves of tracer tests in order to extract the storage rate and cross-sectional area of the storage zone for hyporheic interactions (Lautz et al., 2006). MODPATH, a particle tracking package, can be used to simulate water transport through the hyporheic zone and the resident time distribution of stream

water in the hyporheic zone (Kasahara and Wondzell, 2003; Lautz and Siegel, 2006). MODFLOW is a two- or three-dimensional groundwater model that can be used to simulate water exchange between the stream and an adjacent aquifer (Wondzell and Swanson, 1996b; Kasahara and Wondzell, 2003; Cardenas et al., 2004). MT3D99 has been used for transport simulation, and ZONEBUDGET for the calculations of fluxes in the hyporheic zone (Cardenas et al., 2004). The models themselves and their algorithms will not be discussed here, but model methods and assumptions should be thoroughly investigated before they are implemented. This should ensure that the approximations applied to the process of interest in order to forecast its function are made explicit.

## 2.4 Groundwater Flow Through Peat

The transmission of water through organic soils complicates the standard theories of matrix flow and infiltration that are normally applied in flow calculated with Darcy's Law (Bradley and Berg, 2005). Complications mainly arise due to peat's inherent anisotropy. This section highlights some of the definitions and characteristics of organic peat soils and discusses the current methods for assessing *in situ* hydraulic conductivity estimates.

### 2.4.1 Peat Definition and Discussion

Wetlands in which peat (organic soils) accumulate are broadly defined and definitions can vary from region to region. The term peat indicates a medium that is primarily comprised of decaying and partially decaying organic matter (Mitsch and Gosselink, 1993). In order for a substrate to be classified as peat, the percent organic must be greater than 11%, depending on the clay content (Mitsch and Gosselink, 2000). Peat soils result from high accumulation rates of plant matter compared to slow decomposition rates, which are the result of semi-permanent anaerobic soil conditions found in wetland environments. Peat accumulations can often be very old and very deep (up to 10 m) (Clymo, 2004).



Researchers have distinguished two horizons within peat deposits. The acrotelm is located in the top five to ten centimeters of the deposit (Clymo, 2004). This layer is characterized as containing coarser plant root and biomass material. During dry summers the water table frequently drops below this layer, exposing it to desiccation. The second layer lies below the acrotelm and is referred to as the catotelm (Clymo, 2004). The catotelm is a water-logged, anoxic environment where bacterial activity depletes oxygen and the moisture content and conductivity limits its diffusion (Clymo, 2004). This layer usually remains saturated and there is a marked decrease in the amount of fibrous plant material.

### **2.4.2 Peat Structure Macropores and Piping**

Unlike soils that are deposited and sorted by wind or water peat deposits do not undergo any form of sorting. Thus, peat accumulations are inherently anisotropic (Bradley and Berg, 2005). The physical structure of peat is primarily controlled by the vegetation accumulation from which it is composed. Within the acrotelm is a zone referred to as the “root mat” or “root zone”. This zone is made up of roots, rhizomes and tangled with partially decayed organic matter (Baird et al., 2004). The root zone creates a highly permeable layer with soil structures known as macropores. Macropores are voids or channels within soil structures that preferentially transport water and solutes. Macropores are produced from roots, cracks, and are often the result of worm activity or other burrowing insects (Holden et al., 2001). They can range in size from millimeters to a couple of centimeters or bigger. Macropore flow is an active infiltration and transport process in peat (Quinton et al., 2000). The importance of macropores may decline during wet periods because saturation excess overland flow becomes the dominant runoff process in peatland areas and also because of the collapse of macropores during high intensity precipitation events (Holden et al., 2001). Vegetation type and peat depth seem to be linked to the presence of macropores.

Pipes are similar to macropores in that they provide a preferential conduit for the transport of event water. However, pipes are not always found near the surface

of the peat deposit in the rooting zone; instead they can be found at depth and range from a few centimeters to sizes large enough for a person to crawl into (Holden and Burt, 2002). Pipes are primarily created from eroded preferential flow paths and cavities left from the decay of old tree roots or buried branches. The development of pipes is also correlated with slopes, and areas with little relief are less prone to these larger scale developments (Quinton et al., 2000). Like macropores, pipes may collapse under certain hydrological conditions. They may also become ineffective if the water table is not sufficiently high to induce flow. Pipes, unlike macropores, are large enough to be individually gauged and their contribution to stream flow roughly estimated. Thus, hydraulic conductivity is not normally assessed for these features (Holden and Burt, 2002).

### **2.4.3 The Nature and Methods for Estimating Hydraulic Conductivity in Peat**

The hydraulic conductivity of peat is found to decrease with depth. Aside from this, no other generalizations about conductivity can be made in peat because of the variability from location to location, and even the variability within one deposit (Quinton et al., 2000; Holden et al., 2001; Clymo, 2004; Bradley and Berg, 2005). The decreasing hydraulic conductivity with depth may provide a flow boundary, or transition zone, where water goes from vertical infiltration or percolation through the conductive zone to lateral transport when it reaches the low conductive zone. Seasonal ice layers and the water table can also induce changes in the flow direction within peat. Ponding may take place above these barriers that either completely impede passage to lower layers (seasonal ice) or be located in a less conductive zone (water table) in either situation (Quinton et al., 2000). Identifying the zones of different hydraulic conductivity is essential for understanding these flow diversions and properly estimating the average spatial conductivity of the system.

The individuality of each peat deposit strengthens the need for accurate *in situ* estimates of hydraulic conductivity. Some of the methods most commonly applied in

the estimation of hydraulic conductivity in peat are chloride tracer curves, piezometers in conjunction with rising and falling head tests or pump tests, infiltrometers with a constant head, and lab experimentation on collected soil samples. Estimations made using rising head tests are associated with pore clogging when the slug displaces water into the surrounding peat (Baird et al., 2004). Falling head tests do not have this problem because pores are flushed up into the piezometer tube. Pumping or developing a piezometer before testing can help reduce some of the problems with this method (Baird et al., 2004). Ponding methods, such as estimates made using infiltrometers, may see a decrease in conductivity over successive trials due to macropore destruction (Bradley and Berg, 2005). Lastly, lab analysis of conductivity may incur errors associated with sample damage (compression, micropore collapse etc.) (Baird et al., 2004).

## **2.5 Rationale for Studying Hyporheic Exchange and Biological Processes**

This section will outline the hydrological, chemical, and nutrient importance of hyporheic exchange set within the context of contemporary research, and investigate the rationale for studying the effects of biological processes, such as beaver dams, on hyporheic processes. The goal of this section is to bring together the physical processes regulating hyporheic exchange with the biological activities that may affect these physical processes.

### **2.5.1 The Hydrological Role of Hyporheic Exchange**

Elevated groundwater levels in a saturated hyporheic zone can contribute to, and increase the contribution of, event water in stream discharge (Cey et al., 1998). As a result, a given storm event will see a quicker response to precipitation inputs and peak discharge, which may not be accounted for in flood prediction models, and result in unpredicted damages. This may also be reversed in scenarios where the

groundwater levels within a hyporheic zone are lower than the stream channel and are capable of receiving water inputs from storm elevated stream levels as well as through the infiltration of storm precipitation, thus attenuating the peak discharge and flood damage. The hyporheic zone has an effect on the hydrological routing of ground and surface waters (Lautz et al., 2006). Water in the hyporheic zone may be redirected from its original flow paths under very dry or very wet conditions to either the groundwater or surface water system. These flow paths may be redirected over longer flow lengths and times. For example, surface water directed into a dry groundwater system may not be re-discharged to its originating tributary, but instead carry on in the groundwater system, complicating water management and prediction (Lautz et al., 2006).

### **2.5.2 Chemical Reactions and Nutrient Cycling in The Hyporheic Zone**

The water exchange between the anaerobic groundwater system and the aerobic surface water system creates conditions of tightly spaced oxic and anoxic environments (Triska et al., 1993b; Lautz and Siegel, 2006). These environments have hydrochemical implications for both the stream and groundwater flow systems (Triska et al., 1993b; Lautz and Siegel, 2006). It is within these environments and under these conditions that high rates of nutrient transformations and transport take place such as the denitrification or reduction of nitrate to ammonium in areas with low dissolved oxygen and dissolved organic carbon (Triska et al., 1993b; Lautz and Siegel, 2006). These contrasting environments are also sites for increased redox reactions which feed microbial activity within the active groundwater system (Triska et al., 1993b). These processes help raise the concentrations of solutes and nutrients, and are thus important for plant productivity. Resulting increases in plant productivity can increase bank stability and reduce stream sediment loads. Thus, chemical and nutrient processes taking place in the hyporheic zone are important factors in determining the chemical and biological status of streams (Wondzell and Swanson, 1996b).



**Figure 2.2:** The distribution of beaver in North America (Baker and Hill, 2003)

### 2.5.3 Beaver Dams and Hyporheic Exchange

There are a multitude of biological organisms that affect the function of surface stream systems, but very few have such an obvious and obtrusive effect as the alterations made by beaver and their dams. Dams affect the hydrological behavior of stream systems by ponding water, redirecting flow patterns, and increasing the amount of water available for evaporation (Woo and Waddington, 1990). The habitat range for the two species of beaver (Eurasian beaver *Castor fiber* and the North American beaver *Castor canadensis*) spans throughout the northern forest belt of Canada and the United States (Figure 2.2 to Europe and Asia (Rosell et al., 2005). In North America alone, beaver have a habitat range of approximately 15 million square kilometers. Throughout their range, beaver occupy a number of ecoregions from the subtropical to the subarctic (Rosell et al., 2005). Since their protection in the 1920s, beaver populations have been on the rise due to both natural and

reintroduction processes (Rosell et al., 2005). The population increase of the *C. canadensis* has been greater than that of the *C. fiber*, but both species are continually expanding their range (Rosell et al., 2005). With beaver populations on the rise, their effects become increasingly more important to water management schemes throughout headwater streams in North America and Europe.

To date, it has been shown that beaver dams have a multitude of hydrological effects. Beaver dams have been associated with the creation of upwelling and downwelling zones in stream channels (White, 1990). Dams have been identified as the sources prolonging the inundation of valley flood waters on the floodplain for tens to hundreds of meters downstream (Westbrook et al., 2006). In some areas it has been observed that the redirection of surface water to the hyporheic zone creates deep vertical flowpaths within the valley system (Lautz and Siegel, 2006).

Beaver thus affect the hydrological functioning of stream systems in ways that are not incorporated into the current conceptual framework of stream hydrology. They can influence the seasonal fluctuation of stream stage, and surface and groundwater mixing, creating unique seasonally variable local flow patterns. Research into the hydrological and biological processes described in this section is critical in order to create holistic frameworks for riverine systems in which beaver and their dams are present.

## 2.6 Synthesis

The research being conducted on hyporheic processes and their contributions to greater hydrological processes, both at the surface and subsurface, began in the early-to-mid 1980s. Traditionally, the hyporheic process has been viewed as a subsection of the greater hydrological scheme, and thus has only been studied within that range. However, to study the hyporheic process as a discrete topic of hydrology without considering any biological processes does not reflect the real-world conditions found throughout North American streams, so results obtained to date have limited applicability. This review of the literature has aimed at introducing the importance

and complexity of the hyporheic process in hopes of providing a catalyst for the direction of this thesis.

Hyporheic processes are studied by both ecologists and hydrologists; however, each scientific group focuses on different characteristics of the process. The ecologist may study the effects of hyporheic fluxes on biological populations or vegetation successional patterns, whereas the hydrologist may focus on the influences of flood waves or pool-riffle sequences and how that may affect subsurface flow patterns and sediment loads (White, 1990; Wondzell and Swanson, 1999; Hayashi and Rosenberry, 2002; Cardenas et al., 2004). For example, ecologists have focused on vertical hyporheic flow in order to identify suitable fish spawning habitats. Hydrologists, on the other hand, have focused on lateral hyporheic exchange to better understand bank storage dynamics. However, hydrologists cannot afford to ignore the possible effects of biological processes on the hydrological functioning of a system. These biological effects will need to be incorporated into management strategies, hydrological models, and quality control practices if they prove to be effective at changing the current theoretical or conceptual understanding of subsurface and surface hydrological systems.

Major biological processes such as beaver dams have the potential to act as geological stream or subsurface features. Investigation into this feature's effects on the hyporheic zone is currently taking place in a variety of systems (Woo and Waddington, 1990; Burns and McDonnell, 1998; Lautz and Siegel, 2006; Lautz et al., 2006; Westbrook et al., 2006). However, the temporal and spatial scales along with the methods used in the past to study the effects of beaver dams has been limited in scope. Research is needed on mid-scale (tens of meters), seasonally dynamic hydrological effects of beaver dams on groundwater flow patterns that extends the geographical area of research into the large peat deposits found within Canada, including the hydrologically important and complex areas such as the Canadian Rockies, where research thus far has not been conducted. Both high and low elevation sites are home to peatlands and beavers in the Canadian Rocky Mountains. Many of the wide low relief valleys are populated by willow and poplar species and have the evidence for

the presents of beavers in these areas, thus increasing the need for study in these regions.



## CHAPTER 3

### STUDY OBJECTIVES

Little of the current research on hyporheic exchange has a spatial scope that includes broad, flat floodplains with sustained water tables and large peat accumulations. Most studies have focused on steep headwater streams in arid or temperate environments with strong longitudinal gradients and cobble/gravel beds. Past research has also focused on identifying the channel and geological characteristics of streams that create or sustain hyporheic exchange. Little has been done on the role ecosystem engineers, such as beavers have on hyporheic exchange despite that they are ubiquitous throughout aquatic systems in North America and Europe. The overarching goal of this thesis is thus to examine water exchange between a small stream and a shallow peat aquifer along a reach of a beaver dammed stream.

Objective 1: Determine the influence of small, in-channel beaver dams on vertical hyporheic fluxes.

Hypothesis 1: Beaver dams will enhance vertical infiltration of stream water into the bed environment by increasing stream stage; reducing the stream velocity; and increasing water contact with the bed.

Vertical hyporheic flows may be enhanced due to increases in hydraulic head within the upstream portion of a dammed reach causing surface water to downwell into the streambed (Figure 1.1 zone 1). Lower stream stage and hydraulic head directly downstream of dams can trigger zones of vertical upwelling.

Objective 2: Determine whether small, in-channel beaver dams alter lateral

hyporheic exchange and the pattern of groundwater flow near the channel.

Hypothesis 2: Beaver dams will enhance stream water infiltration into the bank, where it will flow beneath the ground surface in a looping fashion around the dam and back to the stream.

Beaver are expected to create local surface water mounds upstream of dams producing stream to groundwater hydraulic gradients with the nearby saturated bank. The beaver ponds will create a flow reversal, with lateral groundwater recharge at times when base flow discharge conditions would normally be expected. The beaver dams are expected to mound stream water and reduce flow velocity above the dams. Longer contact time should induce flow reversals causing stream water to laterally recharge the bank aquifer.

Enhanced lateral groundwater recharge may discharge downstream of dams because the hydraulic gradient may reverse (high groundwater table juxtaposed to low stream stage) resulting in groundwater discharge to the stream. Such closely spaced hydraulic gradient reversals are expected to create looping flow patterns within the hyporheic zone around the dams.

# CHAPTER 4

## METHODS

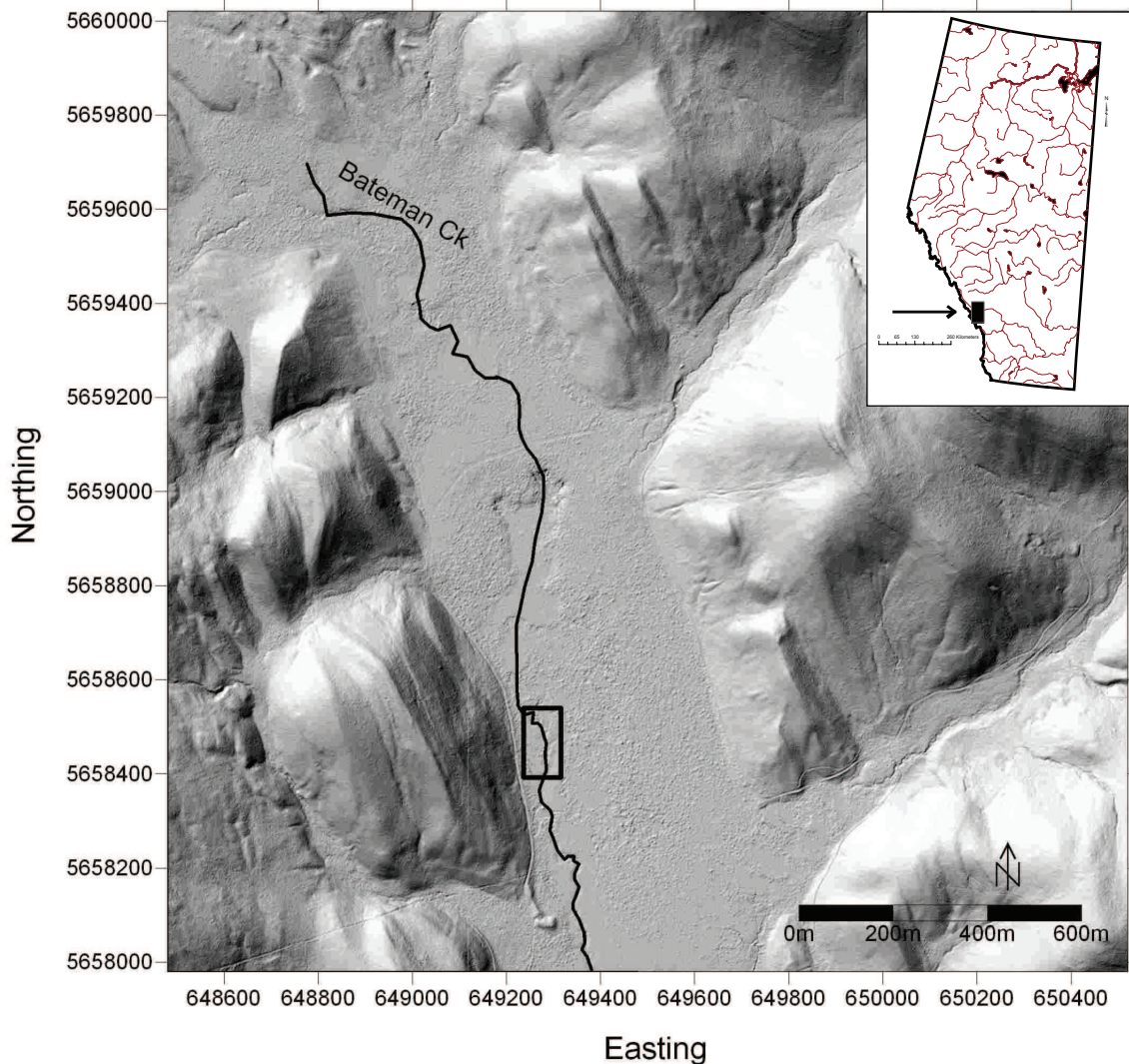
### 4.1 Study Site

The study site, Sibbald Research Basin, is located in the Front Ranges of the Canadian Rocky Mountains, Alberta, Canada (Figure 4.1). Bateman Creek drains the research site and is a second order tributary of Jumpingpound Creek, which feeds the larger Bow River. The valley is 1.8 km long and on average 0.4 km wide and flanked on both sides by foothill mountains that have a maximum elevation of 1653 m with the valley bottom at an elevation of 1480 m.

Mean annual temperature for the Kananaskis Field Station located  $\sim 17$  km from the field site and at an elevation of 1390 m was 3.5 °C from 1975 – 2007. The mean temperature from May to August in 2006 and 2007 was 13.2 °C and 12.7 °C respectively, while the mean temperature for May to August from 1975 to 2007 was 11.8 °C.

The mean total precipitation (rain and snow) for 1975 – 2007 at the Kananaskis Field Stations was 653.0 mm, with 61% falling as rain. Total precipitation for 2006 was 614.0 mm (94% of normal) with 65% falling as rain. In 2007, the total precipitation was 748.0 mm (115% of normal) with 67% falling as rain. The minimum total precipitation value recorded between 1975 – 2007 was 490.3 mm in 1997 and the maximum total precipitation recorded was 951.4 mm in 2005, with 757.4 mm (80%) of rain recorded that year.

The northern part of the valley is dominated by willow (*Salix* spp.), sedges (*Carex* spp.), Sphagnum mosses, dwarfed white spruce (*Picea glauca*) and black spruce (*Picea mariana*). The southern portion of the valley bottom is dominated by



**Figure 4.1:** Location of Sibbald Research Basin in Alberta (inset) and a LiDAR image (2007) of the valley showing Bateman Creek, which drains the valley and the 90 m study reach (rectangle).

sedge (*Carex utriculata*), interspersed with willow. The foothill slopes are comprised of lodgepole pine (*Pinus contorta*), Engelmann spruce (*Picea engelmannii*), white spruce, and black spruce, with pockets of aspen (*Populus tremuloides*).

The research site is located in the southern portion of the valley and is concentrated on a 90 m reach of the stream (rectangle in Figure 4.1). The width of the stream at the study site ranges from 1.2 to 3.0 m. The total relief of the reach is 0.5 m in the down valley direction. The stream banks are predominantly vegetated by *Carex utriculata* and the substrata consists of  $\sim 1.3$  m of peat with lenses of

gravel, sand, and silt throughout. The peat is underlain by light grey to dark grey clays at  $\sim 1.5 - 1.8$  m below the ground surface. The site was flooded because of a previous beaver dam, as observed on an aerial photograph from 1990 showing the dam and flooding extent.

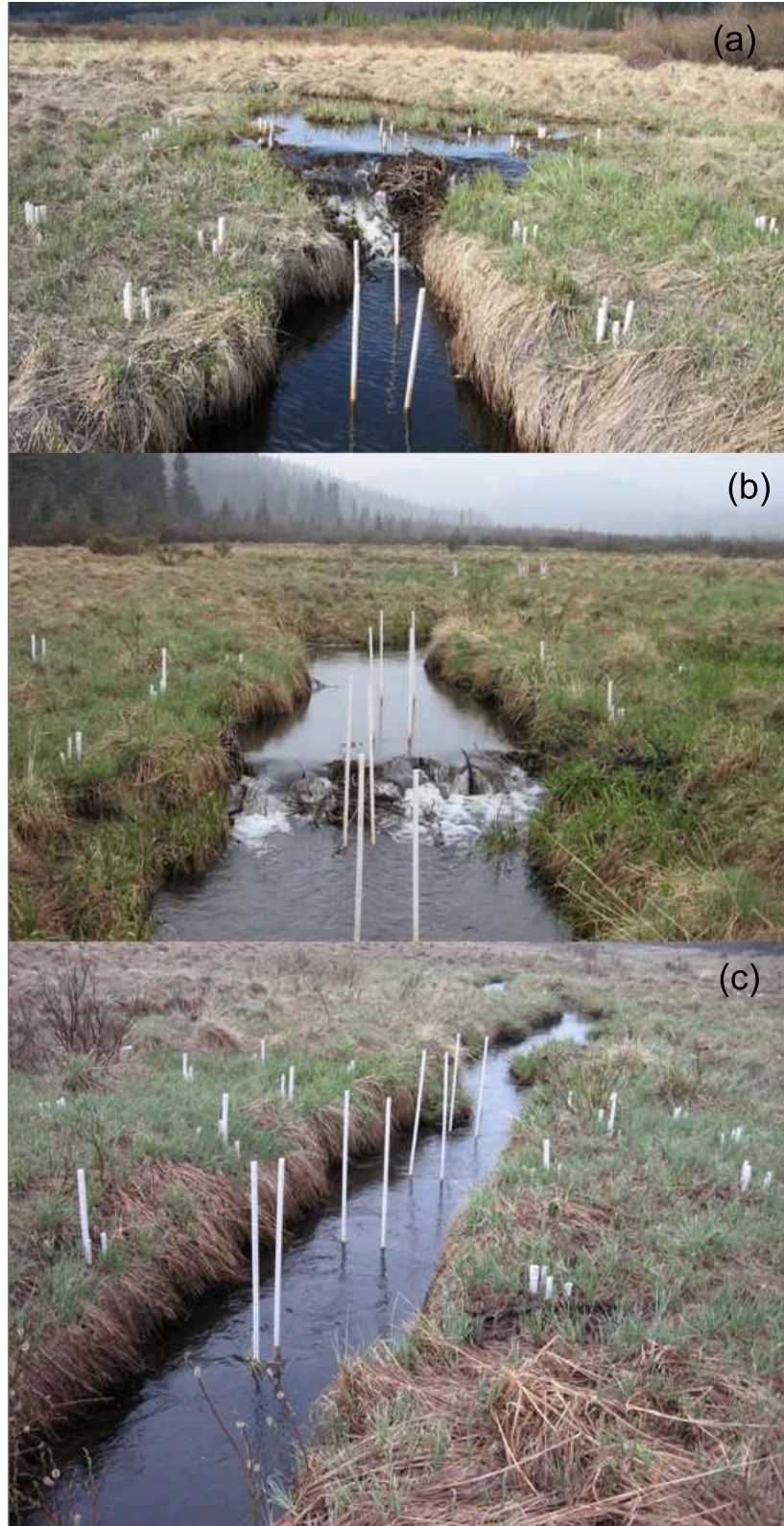
The stream is currently populated by numerous beaver families and dams ( $\sim 14-16$  dams throughout the valley). Seasonal cattle grazing takes place in the fall months within the valley and on the flanking slopes. The natural fauna includes moose, elk, black bears, grizzly bears, cougars, whitetail deer, waterfowl, cranes, and others.

The study reach consists of two sections dammed by beaver (north and south dams) and a third section that remained undammed (reference) throughout the project duration (Figure 4.2). A third dam (Hoover) was constructed some time around 11 July 2007, and was located between the south dam and the reference section. All dams were constructed of willow branches and mud. The north dam was 1.05 m high, 1.71 m wide and likely the oldest dam because it had scoured out a large pool just downstream, the south dam was 0.67 m high, 1.97 m wide and appeared to be newer than the north dam, and Hoover dam was 1.00 m high and 2.70 m wide during the summer of 2007. Instream beaver dams are dams that impede stream flow, raising the water level upstream and lowering the water level downstream, but allow all stream water to remain in the confines of the stream banks; i.e. they do not create ponds. The north dam remained in good repair and similar condition as stated above throughout the summers of 2006 and 2007. The south dam was twice in disrepair starting around the 9 June 2006, and completely flooded out on the 16 June 2006, during two large precipitation events. The dam was completely repaired by the 13 August 2006, and remained in good condition until the construction of the newer Hoover dam flooded it out permanently.

## 4.2 Field Methods

The study design consisted of twelve transects running perpendicular to the stream (Figures 4.2 and 4.3). Each transect contained five piezometer nests consisting of





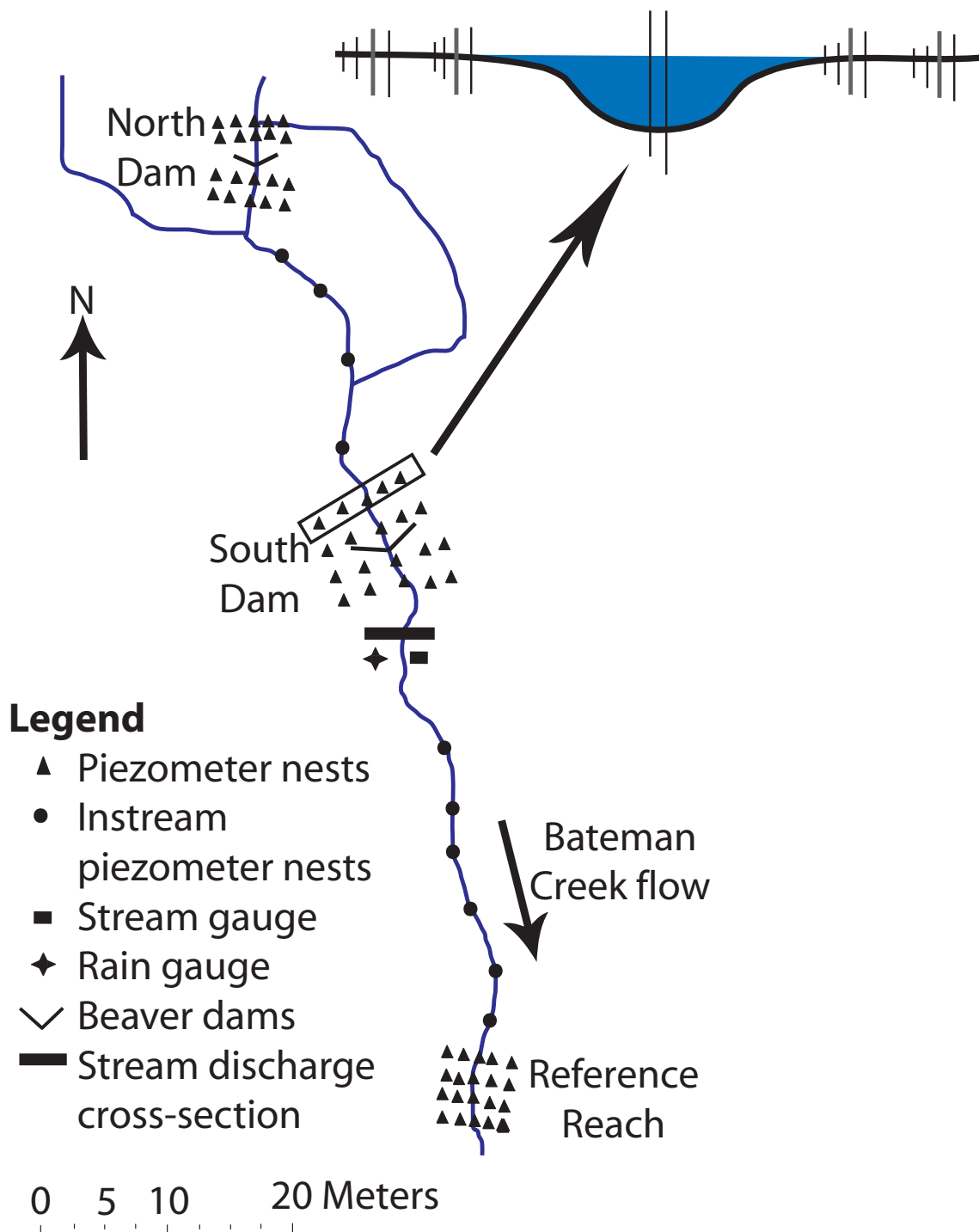
**Figure 4.2:** (a) The north dam (b) the south dam, and (c) the reference section.

three piezometers and one well. Of these five nests, two were placed on each stream bank and one within the channel. The transects were installed so that there were two transects up, and two downstream of the north and south instream dams. The remaining four transects were placed at approximately equal distances ( $\sim 5$  m) in the undammed reference section. Bank nests were situated so that one nest was installed as close to the bank edge as possible and a second was placed  $\sim 1 - 2$  m into the riparian area. The in-channel nests were placed in the thalweg. An additional ten instream piezometer nests were installed in the stream's thalweg between the transect groups in 2007 with four located between the north and south dams and six between the south dam and the reference section.

The bank wells were made of 2.54 cm (inner diameter) PVC pipe that was 1.5 m long, slotted every 5 cm, and capped at the end so that sediment did not enter the well during installation. The wells were hand augered into the ground, leaving a 30 cm above ground 'stick-up' (SU). The instream wells were simulated by measuring the stream stage on the outside of the 'A' piezometers.

The bank piezometers were made of 1.27 cm (inner diameter) PVC pipe, left open at the bottom, and installed with a hand auger in nests of three at 50 cm (A), 75 cm (B), and 140 cm (C), or 75 cm (A), 100 cm (B), and 140 cm (C) below the ground surface. Instream piezometers were made of the same material and driven into the bed to depths of 60 cm (A) and 85 cm (B) by hand. Only the instream piezometer pipes were capped at the bottom and the bottom 10 cm were slotted: the slots were screened with mesh drywall tape to avoid clogging during installation. The SUs for these piezometers were well above the stream bank surface to prevent flood water from entering.

Depth to water in wells and piezometers was measured using a voltmeter wired to a length of graded cable that allowed an electrical current to pass once the open end of the cable encountered water. Stream stage was measured in two ways: it was measured in each of the instream wells using a tape measure, and it was continuously measured (15 min intervals) in the middle of the reach using an OTT Thalimedes Shaft Encoder Level Sensor. Piezometer, well and instream stage water



**Figure 4.3:** The study reach showing the location of piezometer nests, the stream gauge, and the rain gauges (TB3 tipping bucket, Hobo tipping bucket, and manual rain gauge).



levels were sampled at a proximately weekly intervals between June and August in 2006 and 2007, totaling 8 sample dates per year. Precipitation was monitored using an automated tipping bucket rain gauge (Model TB3, Hydrological Services). The northings, eastings and elevation for all instruments and beaver dams were surveyed using a total station.

### 4.3 Data Analysis

Seasonal mean stream stage was separately calculated for the periods before the Hoover dam and after, which gave some indication of the changes this large beaver dam induced. A stage discharge relationship was used to establish rating curves (Figure A.1) that allowed for the estimation of discharge from continuous stream stage collected by a float pot recorder. Separate rating curves were produced for the 2006 and 2007 field seasons because different velocity meters were used for estimating discharge. Daily cumulative values of precipitation were used to determine whether changes in stream stage were affected by beaver activity or by rain events.

Estimates of hydraulic conductivity were completed using falling head tests in all piezometer nests. Changes in head were induced by quickly pouring water in the well or piezometer until it was full. Drops in head were monitored using either an automated pressure transducer (PT2X, Instrumentation Northwest Inc.) put in place before the raise in head, or by manually measuring water levels at sites where water levels recovered very slowly (more than a few hours). Values of hydraulic conductivity were calculated using the Hvorslev method (?),

$$K = \frac{r_c^2 \ln(\frac{R_e}{r_s})}{2L_e T_0} \quad (4.1)$$

where  $r_c$  (cm) is the well radius,  $R_e$  (cm) is the effective radius of the slug test,  $r_s$  (cm) is the radius of the well screen,  $L_e$  (cm) is the length of the well screen, and  $T_0$  (s) is the time at which a normalized head of 0.368 is obtained. This method requires an estimate of  $R_e$ , which is an empirical parameter that is a function of  $\alpha$ , a dimensionless storage parameter (?). The storage parameter is a function of the

radius of the well screen and well, as well as the specific storage,  $S_s$  ( $\text{cm}^{-1}$ ):

$$\alpha = \frac{r_s^2 S_s}{r_c^2}. \quad (4.2)$$

In the Hvorslev method, values of  $R_e$  are typically either the well screen length or 200 times the effective radius of the well screen.  $R_e$  was approximated to the screen length of the well or piezometer.  $S_s$  ( $\text{cm}^{-1}$ ) is defined to be the volume of water that a unit volume of aquifer beneath the area released in response to a unit decrease in head (Dingman, 2002).

Stream stage measurements at the instream wells were converted to height above streambed by subtracting the water level measured from the total length of the piezometer SU above the streambed surface. Mean values for each instream well were computed for 2006 and 2007, then plotted longitudinally above the surveyed streambed.

Specific discharges were calculated from a modified Darcy's law,

$$q = K \frac{\Delta h}{\Delta l} \quad (4.3)$$

for which  $K$  (m/s) is the estimated hydraulic conductivity and is denoted as a positive value so that discharges from the stream are defined to be negative and discharges into the stream are then positive. Also in equation (4.3)  $\Delta h$  (m)/ $\Delta l$  (m) is the hydraulic gradient. Vertical hydraulic gradients (VHG) between the streambed and underlying aquifer were estimated using the methods outlined in Baxter et al. (2003),

$$VHG = \frac{\Delta h}{\Delta l} \quad (4.4)$$

where  $\Delta h$  (m) is the difference in head between the piezometer of interest and stream stage, and  $\Delta l$  (m) is the distance from the streambed surface to the first slot in the piezometer's screened length. Lateral gradients were calculated by dividing the difference in hydraulic head between the nearest bank piezometer and the adjacent stream stage head ( $\Delta h$ ) (m) by the horizontal distance between the bank piezometer and the instream well ( $\Delta l$ ) (m).

Vertical and lateral volumetric discharges,  $Q$  (L/s), were the product of  $q$  (m/s) and area ( $\text{m}^2$ ),  $A = L \times W$ , over which the flux is calculated. For the vertical

flux,  $L$  (m) was estimated as the sum of half the distance between the piezometer of interest and its nearest upstream neighbour and half the distance to its nearest downstream neighbour, while  $W$  (m) was estimated as the distance from stream bank to stream bank at that piezometer. The value of  $L$  (m) for lateral fluxes was determined as the depth from the top of the stream bank to the bottom of the streambed,  $W$  (m) was the sum of half the distance between the bank piezometer in question and the one immediately to the north and half the distance to the closest piezometer to the south.

Flow nets were created by kriging the point measurements of the wells mean hydraulic head values sampled for each of the 2006 and 2007 eight sample dates. Kriging was performed using Surfer ver. 8 (Golden Software Ltd.) because it is an unbiased linear estimator of interpolated values. A linear variogram model was used for all plots.

Standard errors were calculated for the 2006 and 2007 mean VHGs and fluxes using the formula

$$SE = \frac{S}{\sqrt{N}}, \quad (4.5)$$

where  $N$  is the number of data points and the standard deviation,  $S$ , was computed as

$$S = \sqrt{\frac{\sum (x_i - \bar{x})^2}{N - 1}}, \quad (4.6)$$

where  $x$  is the sample value and  $\bar{x}$  is the seasonal mean value (Ebdon, 1985).

# CHAPTER 5

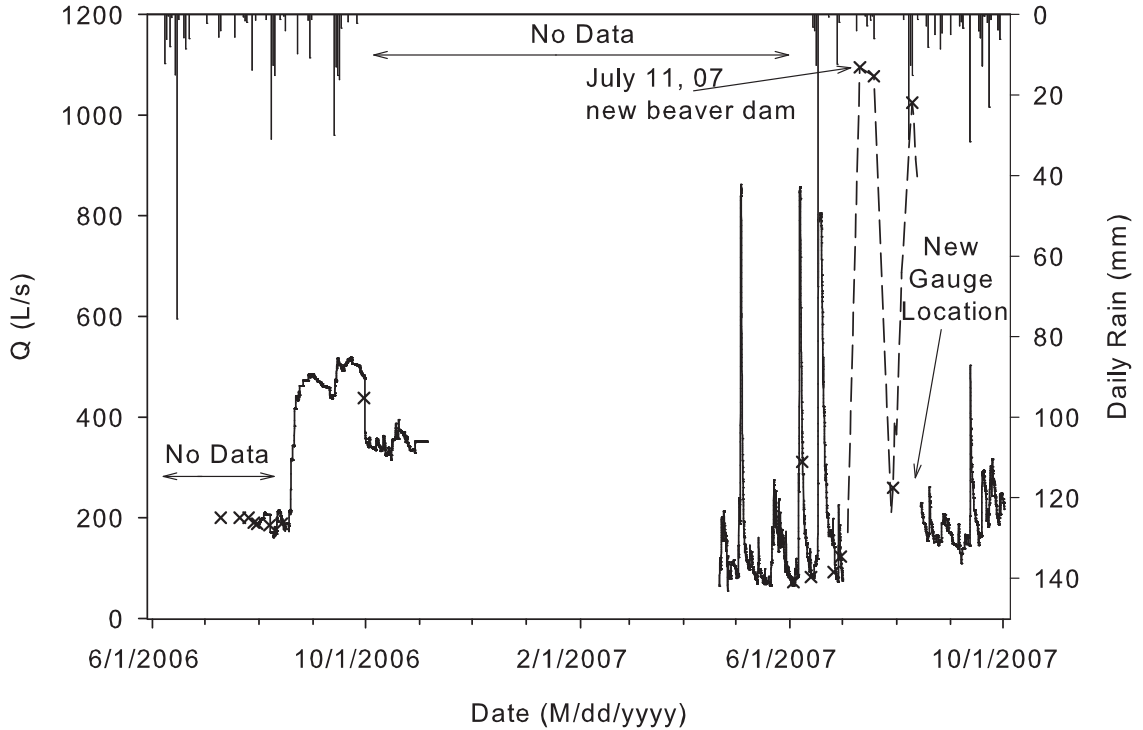
## RESULTS

### 5.1 Stream Discharge and Rainfall

A positive correspondence between rain events and stream stage was found for 2006 and the first half of 2007 (Figure 5.1). For example, all rainstorms  $> 20$  mm produced sizable increases in streamflow. After 11 July 2007, the new Hoover dam raised the stream stage upstream, controlling the level of the stilling well until it was moved downstream of the Hoover dam on 14 August 2007. Stream discharge in the fall of 2006 was more stable than in 2007. In 2007, the stream was much more responsive to small ( $<20$  mm) precipitation events.

### 5.2 Hydraulic Conductivities

Frequency plots of saturated hydraulic conductivities ( $K$ ) for the 48 bank wells show values in the  $10^{-5} - 10^{-6}$  m/s range (Figure 5.2). Of the 48 wells, 24 were located in peat, 11 in gravel, 10 in silt and only 3 wells were classified as being completed in clay. The frequency plot of  $K$  for the 144 bank and 44 instream piezometers indicates that most piezometers have conductivities that are in the  $10^{-6}$  to  $10^{-8}$  m/s range, which Fetter (2001) classified as a silt. Estimates of saturated hydraulic conductivity were conducted on all 48 bank wells and a total of 188 piezometers only one piezometer was excluded from further analysis because the slug test was suspect. During field estimates of saturated hydraulic conductivity well and piezometer heads recovered on average in 1 to 3 hours.

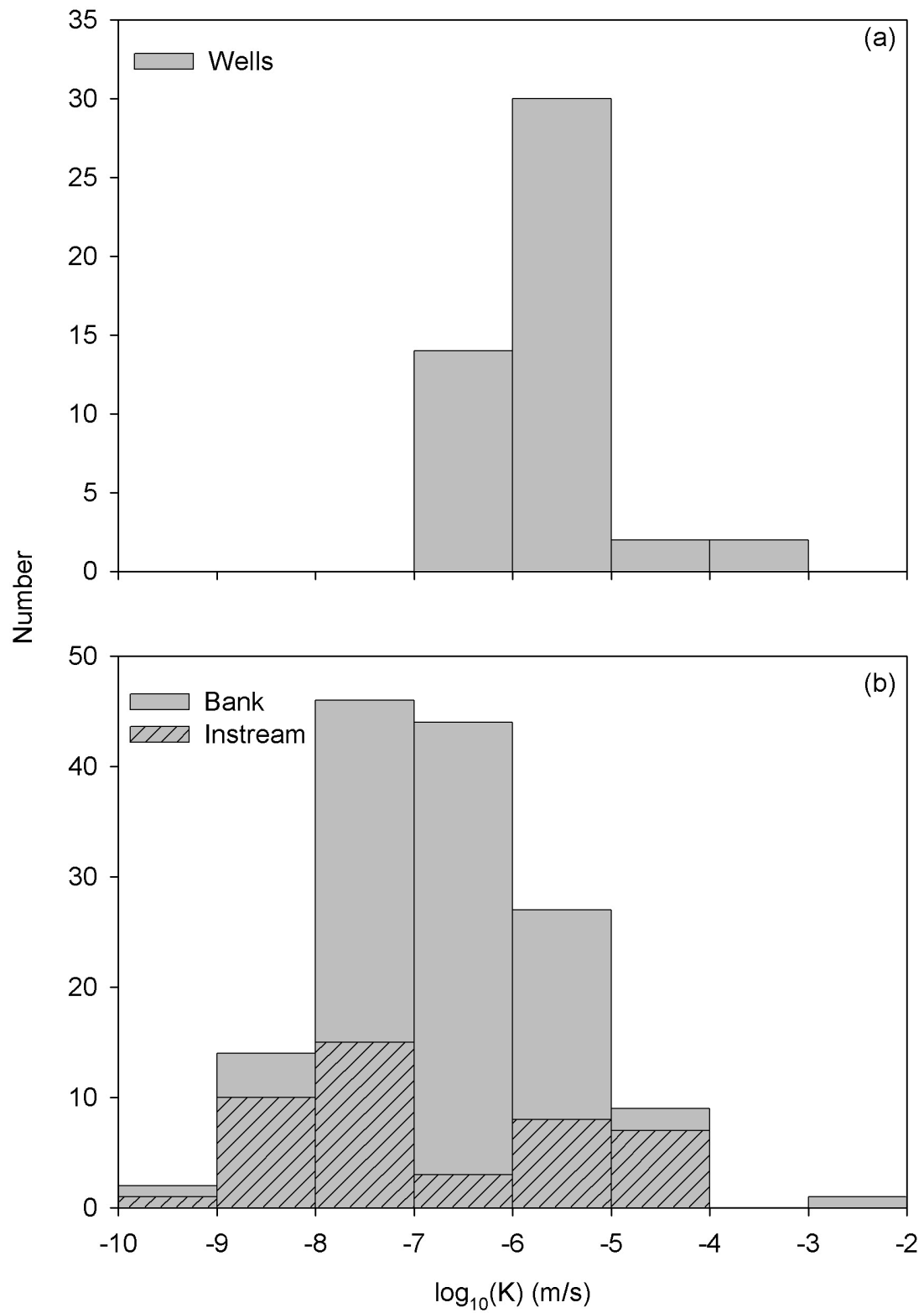


**Figure 5.1:** Stream discharge ( $Q$ ) and total daily rain for the summers of 2006 and 2007. Crosses indicated days when piezometer nests water levels were monitored and dashed discharge indicate calculation from point stage data.

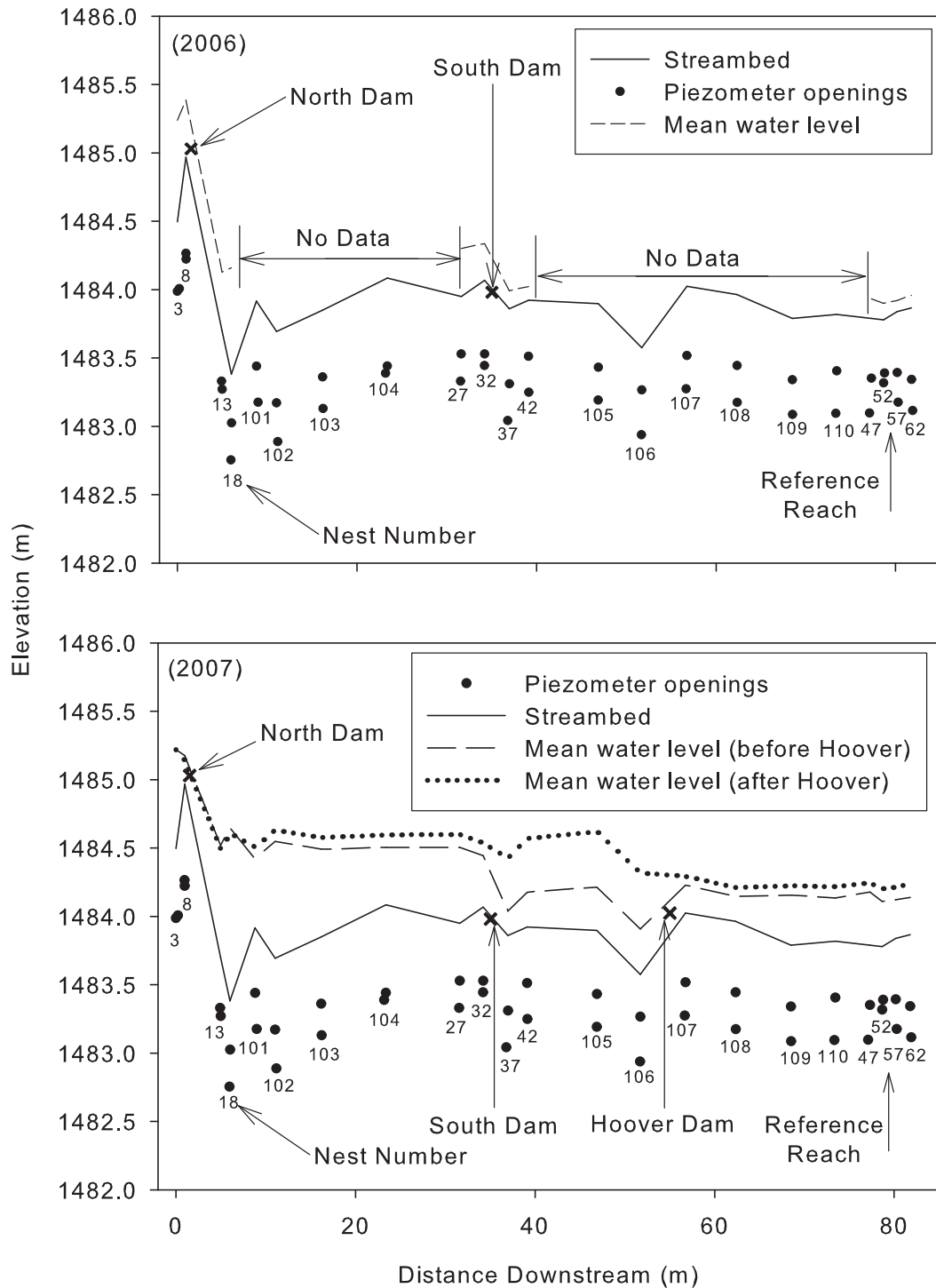
### 5.3 Vertical Hyporheic Exchange

The longitudinal stream profile for 2006 (Figure 5.3) shows the seasonal average water table at each instream piezometer nest for the three study areas. The effects of the south dam are clearly seen in the 30 cm higher stream stage observed upstream of the dam. The stream stage downstream of the north dam was higher likely due to the cutting action of the water flowing over the dam, which created a large scour hole in the streambed immediately downstream of the dam.

Longitudinal stream profiles for 2006 and the first part of 2007 show higher water levels for 2007 in the most northern and southern portions of the reach (Figure 5.3). Following the creation of the Hoover dam, a small increase in stream stage near the north dam was observed. There was a large increase in stream stage observed



**Figure 5.2:** a) Frequency graph showing hydraulic conductivity ( $K$ ) for wells and b) for the bank and instream piezometers.



**Figure 5.3:** Longitudinal stream profiles showing seasonal mean stream stage for 2006 and 2007 as well as location of instream piezometer openings.

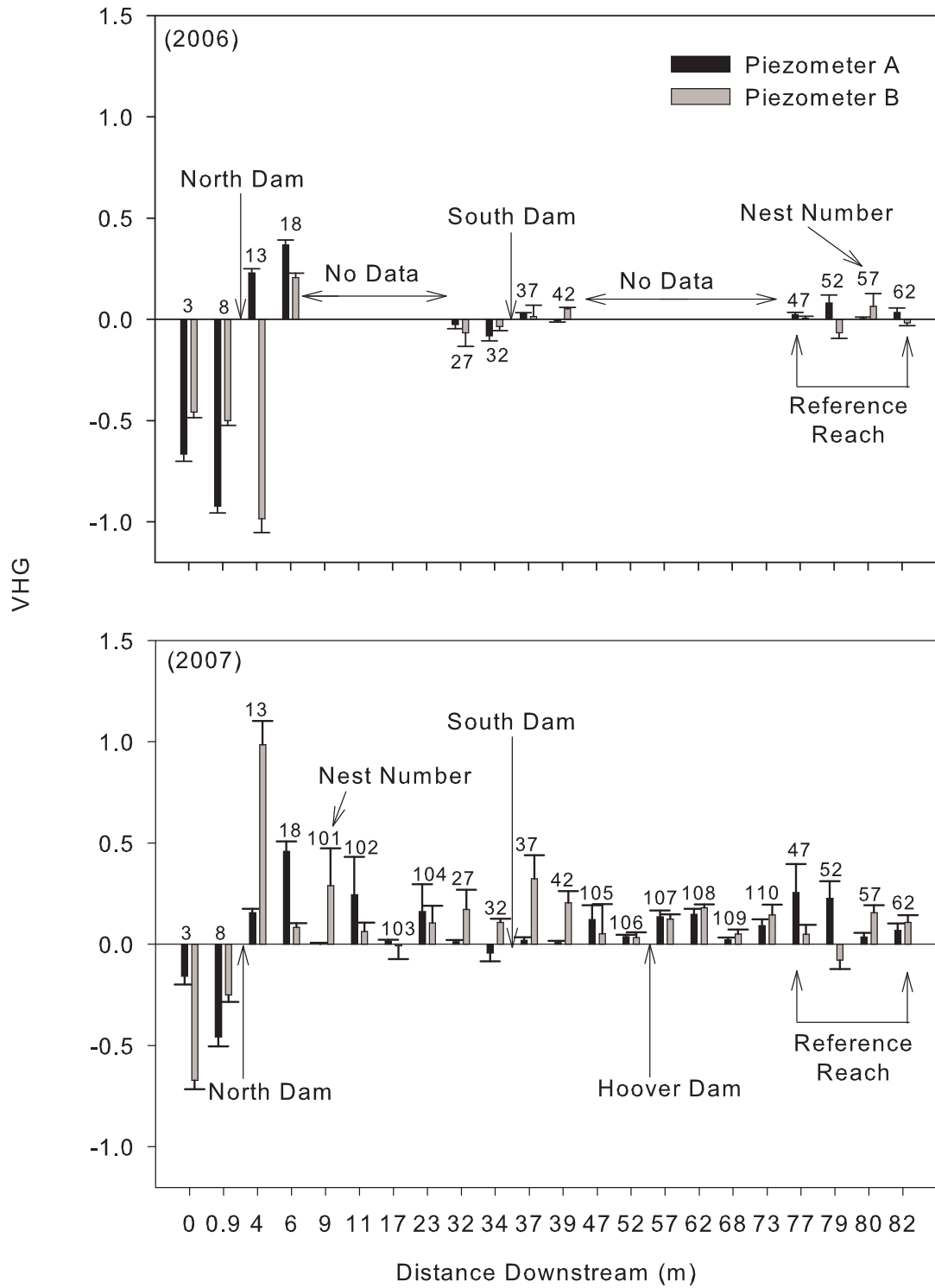
between the south dam and the new Hoover dam. Little difference in stage was seen downstream of the Hoover dam.

Prior to the creation of the Hoover dam, stream stage in both 2006 and 2007 upstream of the south dam was much higher than it was downstream. After the construction of the Hoover dam, a stream stage increase of approximately 50 cm was measured directly upstream of the dam, flooding out the south dam. As in 2006, stream stage downstream of the north dam was higher than immediately above it, due to the presence of a deep scour pool in the bed.

The 2006 mean VHG for the instream piezometers show that gradients are larger around the north than in the undammed reach (Figure 5.4 (2006)). Upwelling and downwelling patterns are formed where water pools upstream of a dam, creating a greater head that forces water into the streambed. Downstream of the dam, this water then recharges back to the stream from the groundwater due to lower hydraulic head. Upwelling and downwelling patterns were found up and down stream of the north and south dams (Figure 5.4 (2006)). The exception to this trend is piezometer 13B, which showed a large downwelling gradient downstream of the north dam. Figure B.1 (2006) shows subsurface head differences for instream piezometer A and B. This graph shows no flow between the shallow A piezometer and the deeper B piezometer in the first piezometer nest upstream of the north dam. The head values of the second piezometer nest upstream of the north dam indicates that subsurface flow is directed upward from the deeper B piezometer to the shallower A piezometer. Subsurface head values between A and B in the second piezometer nest downstream of the north dam identifies a downward flow from A to B.

The 2007 mean VHG for instream piezometers only show the upwelling/downwelling trend around the north dam (Figure 5.4 (2007)). Flow nets (Figures 5.11, 5.10, 5.12, and 5.13) combined with measures of VHG indicate that the remainder of the stream reach is weakly gaining. In 2007, piezometer 13B has an upwelling vertical gradient of equal magnitude to its downwelling gradient measured in 2006 (Figure 5.4 (2006) and (2007)). Subsurface head values for the A and B piezometers for 2007 (Figure B.1) in the first piezometer nest upstream of the north dam indicates





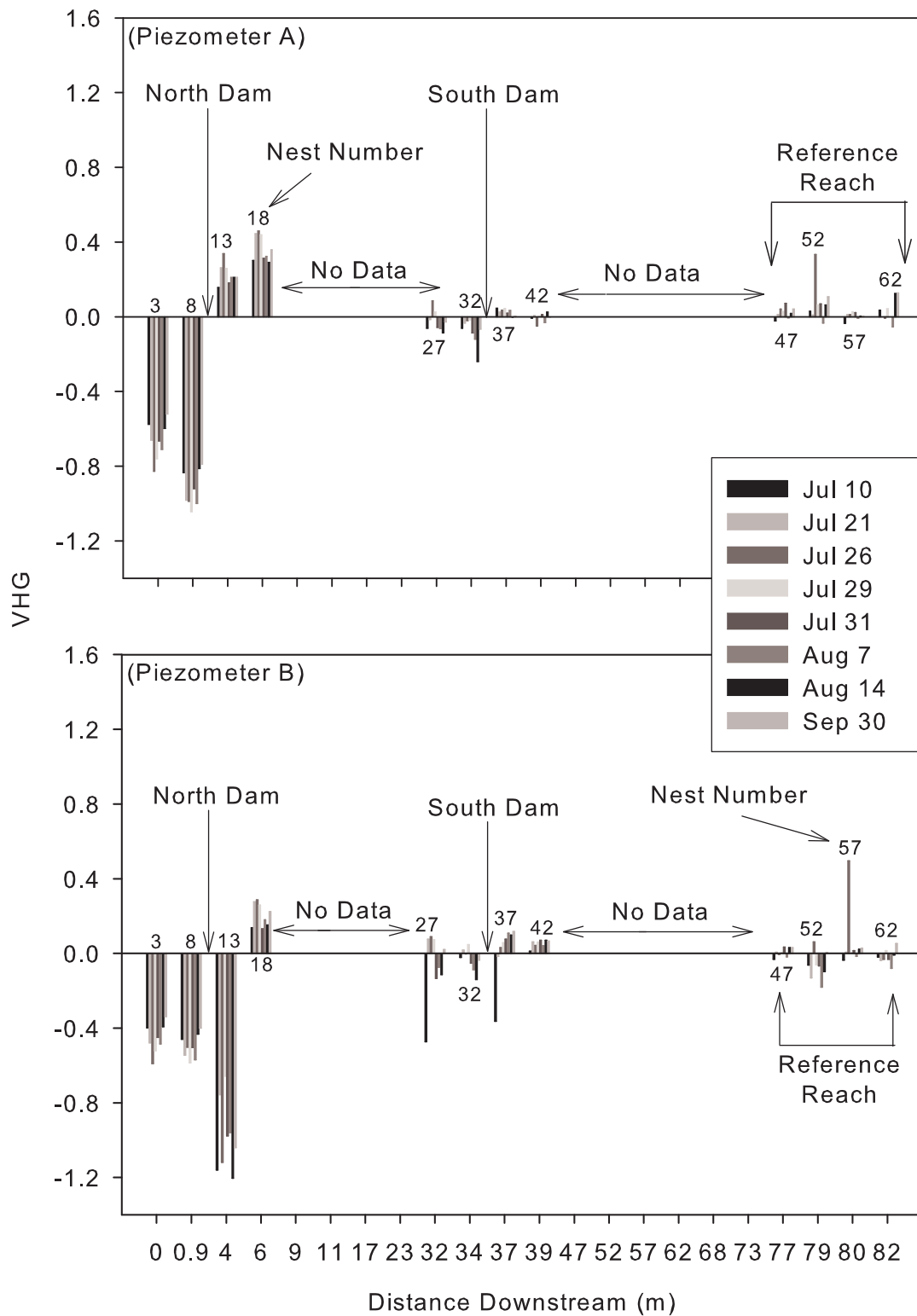
**Figure 5.4:** Seasonal average instream vertical hydraulic gradients with standard error bars for 2006, 2007 piezometer A and B.

downward flow from A to B. The second piezometer nest upstream of the north dam, however, shows a similar upward flow from B to A as 2006 head values. The first piezometer nest downstream of the north dam has reversed its flow from the 2006 pattern and now shows upward flow from B to A. The second piezometer nest downstream of the north dam shows the same downward subsurface flow as 2006.

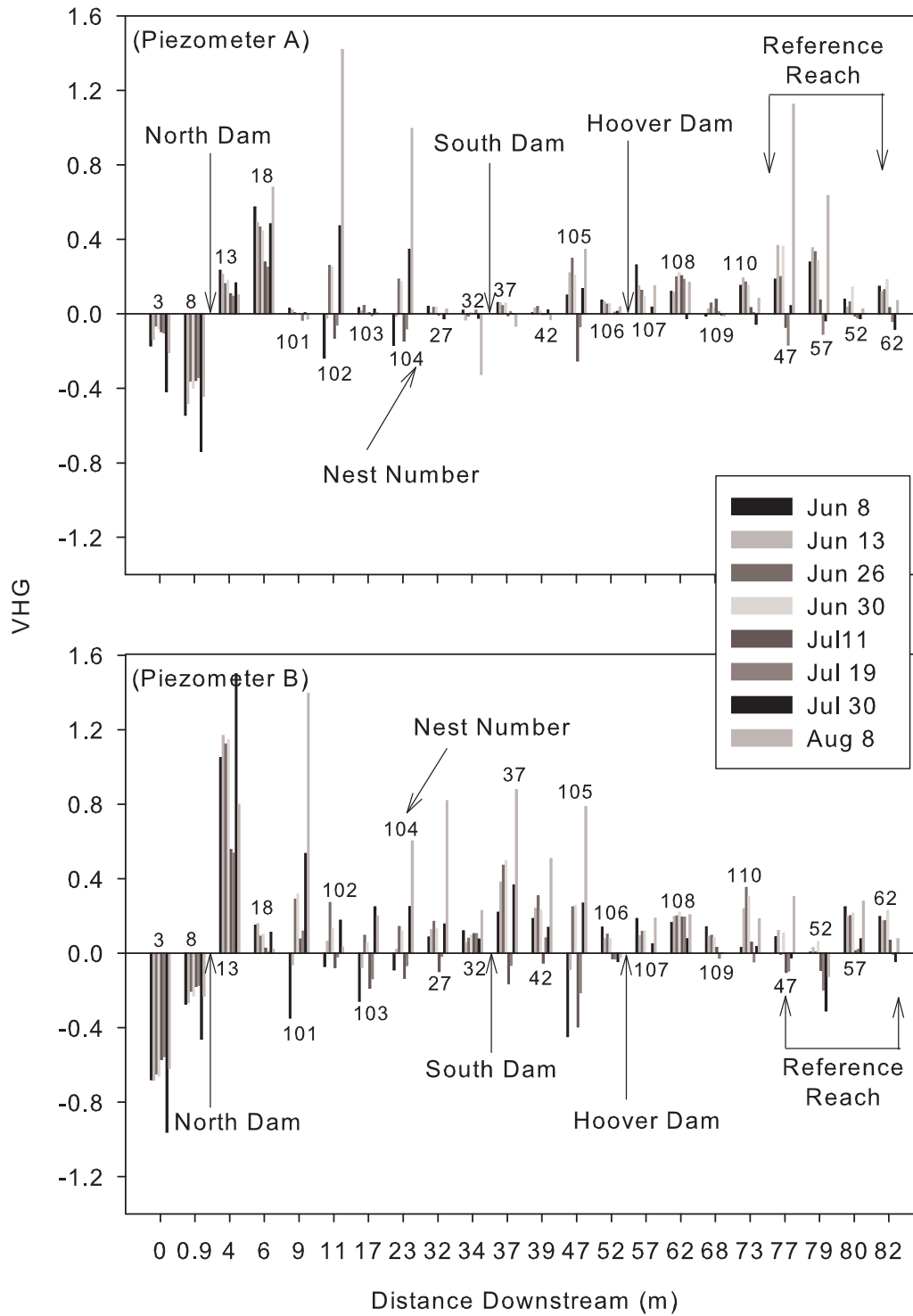
VHG for every date sampled in 2006 show that the A piezometers directly upstream and downstream of the north dam are consistently downward and upward, respectively, irrespective of stream stage 5.5. Similar results were also observed in piezometers 32A and 37A, located directly upstream and downstream of the south dam (Figure 5.5). The remaining A piezometers had mostly upward gradients throughout the summer of 2006. A similar trend in VHGs for the B piezometers was observed, except that there was no looping of flow under the south dam (Figure 5.5). In 2007, VHGs showed flow looping under the north dam in piezometer A and B each sampling day. No other trends in either A or B piezometers were found (Figure 5.6).

Vertical discharge in 2006 shows downwelling upstream of the north dam with no upwelling downstream (Figure 5.7 (2006)). There is upwelling and downwelling upstream and downstream of the south dam in the shallow A piezometers. However, flux was low in the reference section (Figure 5.7).

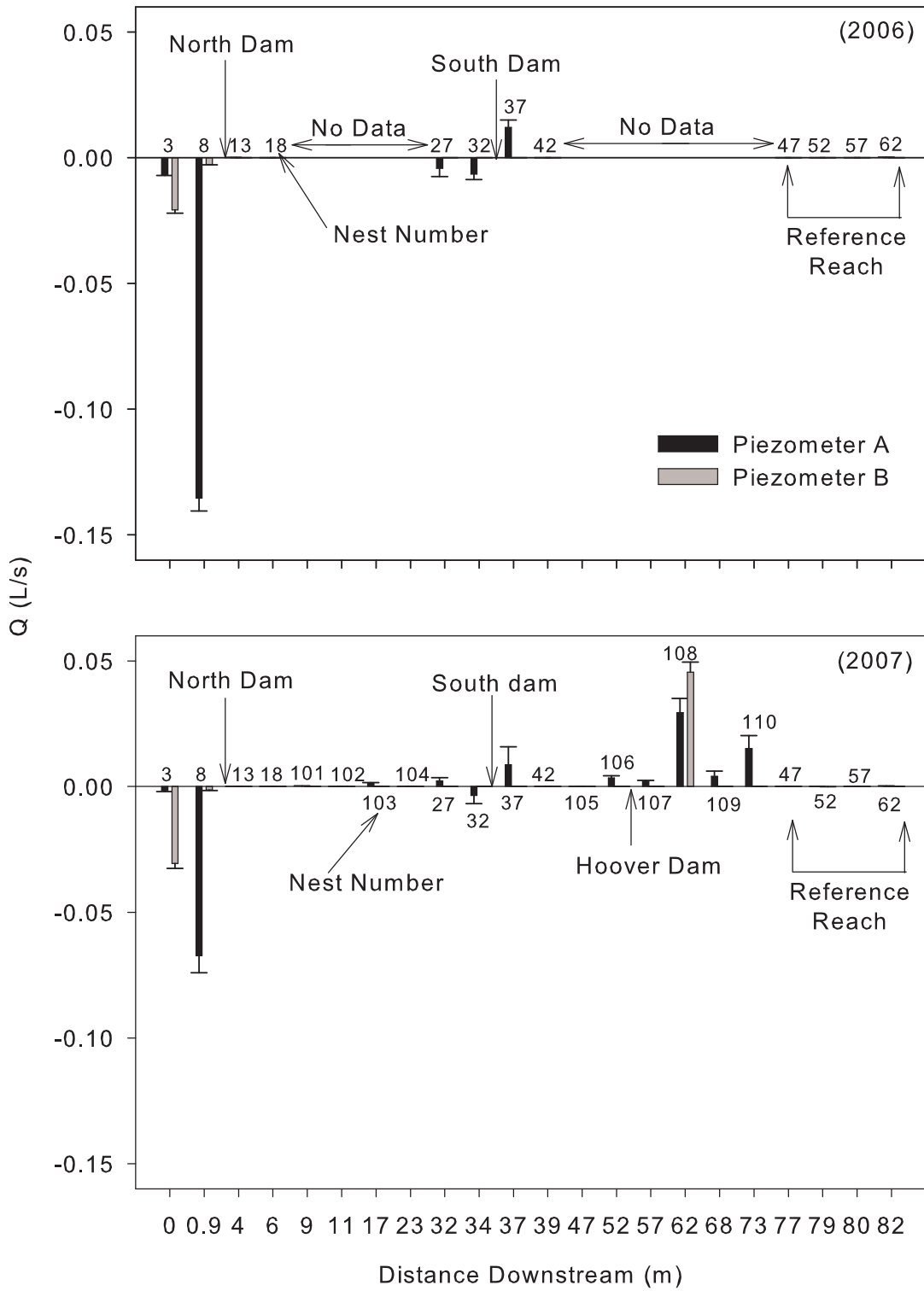
Despite fairly large VHGs around the north dam, small volumes of water are lost to the underlining groundwater system upstream of it with a maximum mean of 0.14 L/s (nest 8) in 2006 5.7. This flux represents  $\sim 0.1\%$  of the mean stream discharge. However, the water is not returning to the stream downstream of the dam, as vertical fluxes there were near zero. Some shallow vertical hyporheic exchange did occur upstream and downstream of the south dam in 2006, but values were  $< 1$  L/s. Between the south dam and reference reach the stream gained water from the groundwater system (piezometer nests 106 through 110). The largest gain ( $\sim 0.4$  L/s) was observed 9 m downstream of the Hoover dam. The reference reach itself did not gain water from, or lose water to the groundwater system.



**Figure 5.5:** Instream VHG for every day sampled in 2006 for piezometers A and B.



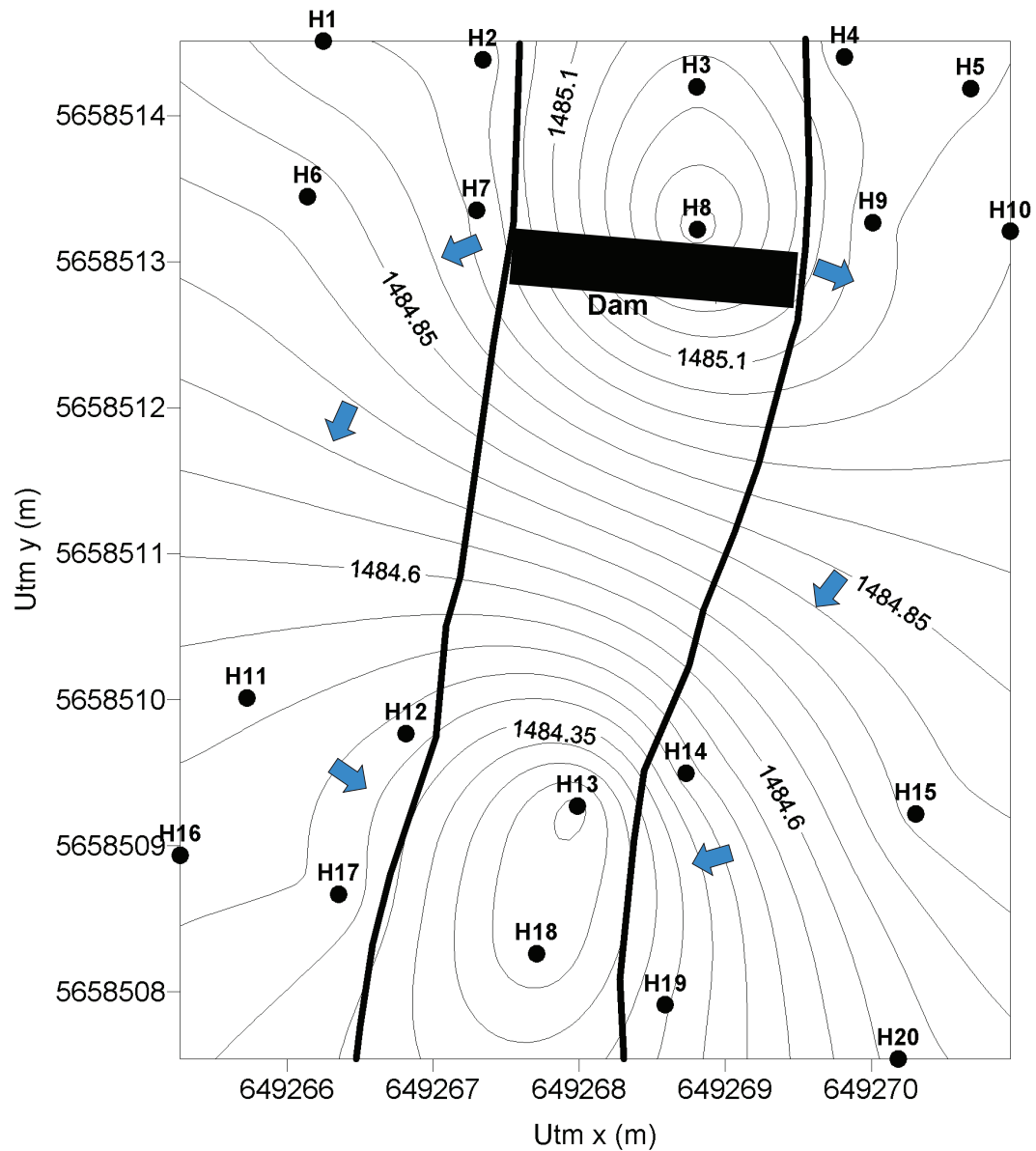
**Figure 5.6:** Instream VHG for every day sampled in 2007 for piezometers A and B.



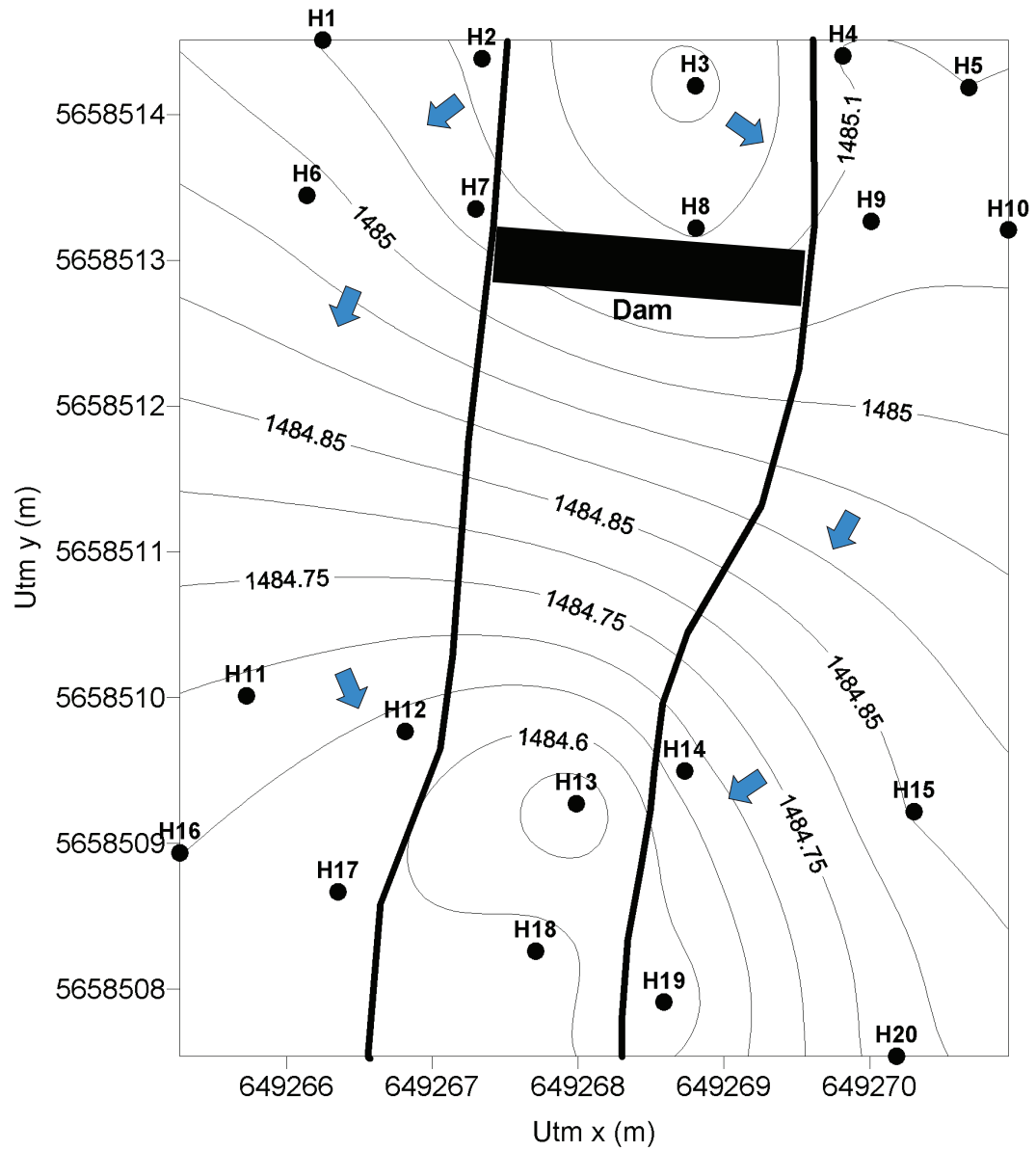
**Figure 5.7:** Seasonal mean discharge with standard error bars for piezometers A and B in 2006 (top panel) and 2007 (bottom panel).

## 5.4 Lateral Hyporheic Exchange

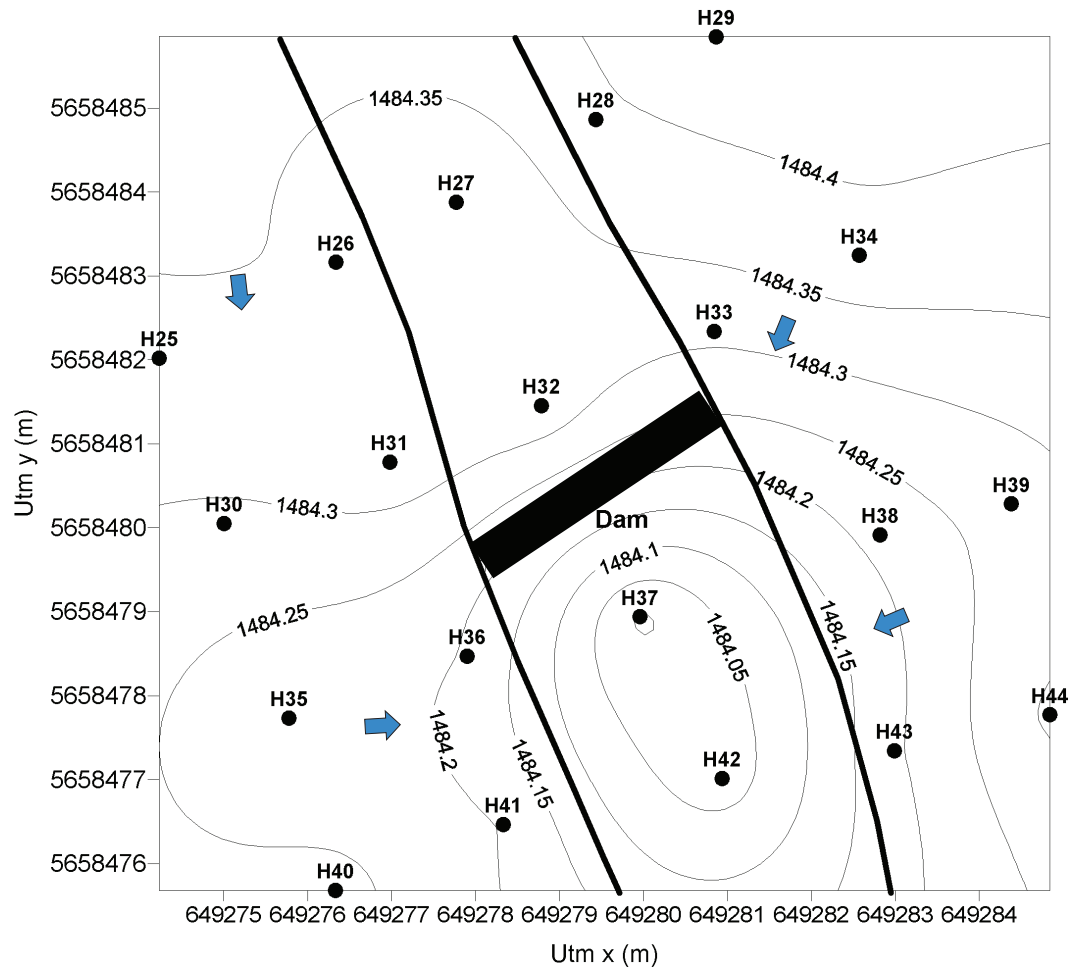
The flow nets for the north dam portion of the study reach show a similar pattern of flow in 2006 and 2007 (Figures 5.8 and 5.9). The flow nets indicate a lateral looping pattern of water around the dam, where there was surface water recharging the bank upstream of the dam and groundwater discharged back into the stream downstream of the dam.



**Figure 5.8:** North dam plan view flow net of mean well hydraulic head (m) for all eight sample dates in 2006

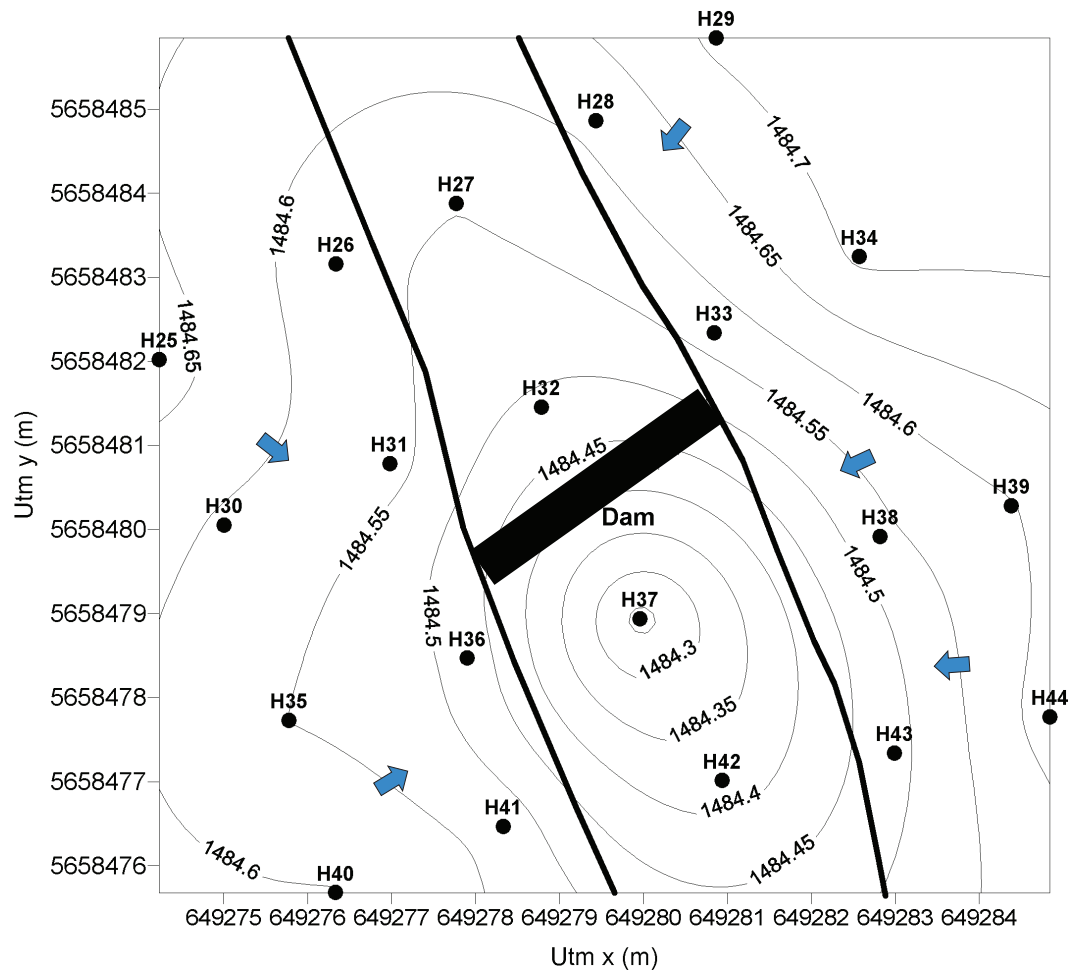


**Figure 5.9:** North dam plan view flow net of mean well hydraulic head (m) for all eight sample date in 2007



**Figure 5.10:** South dam plan view flow net of mean well hydraulic head (m) for all eight sample dates in 2006





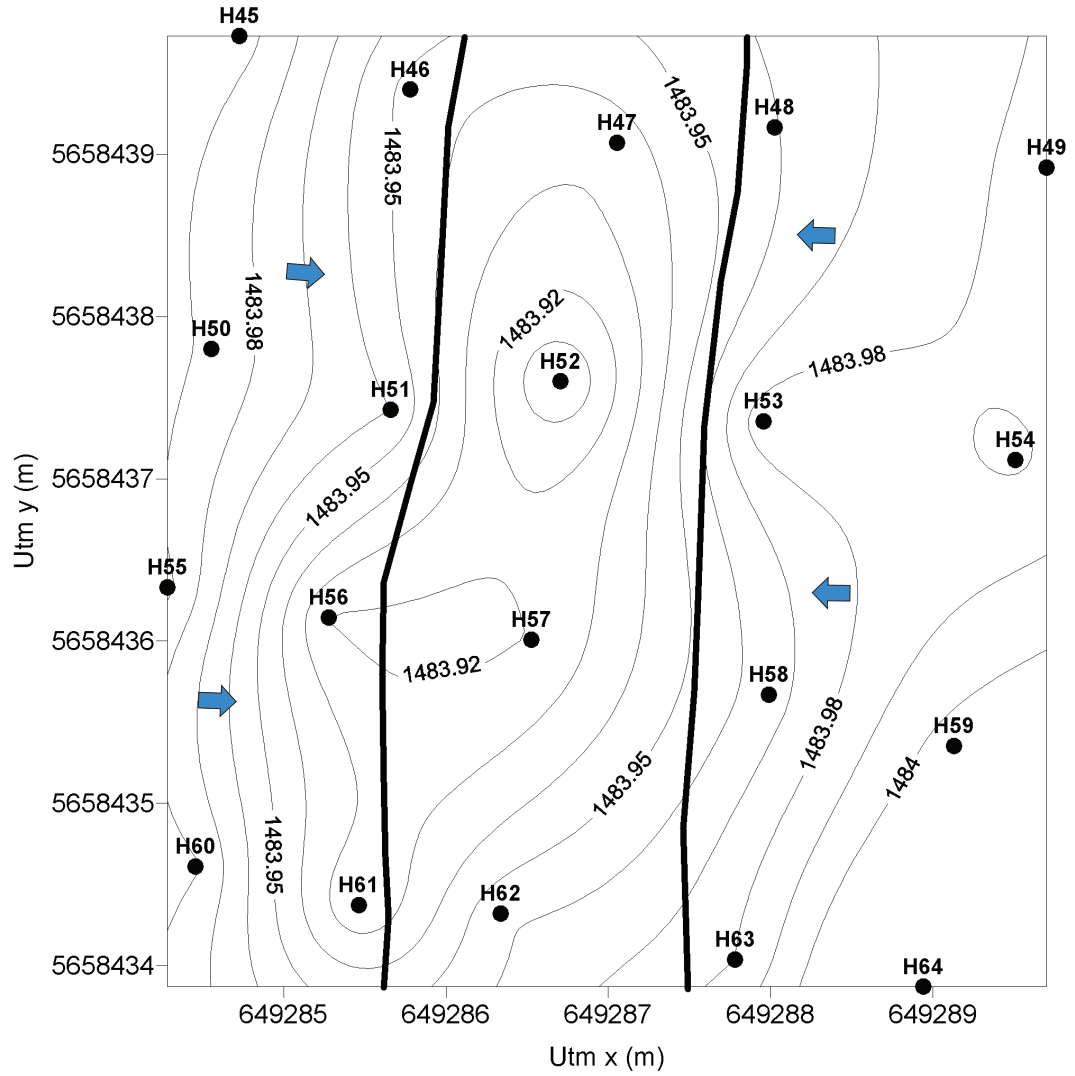
**Figure 5.11:** South dam plan view flow net of mean well hydraulic head (m) for all eight sample dates in 2007

Flow nets for the south dam in 2006 and 2007 showed similar patterns of flow (Figures 5.10 and 5.11). The highest head values were found upstream of the dam, but little flow was directed out of the stream and into the bank because heads were similar in the banks and stream channel. Groundwater flow just downstream of the dam was directed toward the stream from the adjacent floodplain. In 2006, stronger groundwater-to-stream water gradients were measured on the east floodplain than the west floodplain. In 2007, when the south dam was in disrepair, there was more transverse flow of water from the east floodplain toward the stream both above and below the south dam (Figure 5.11). The equipotential lines are aligned with topographic contours in this section of the study reach (Figure C.2).

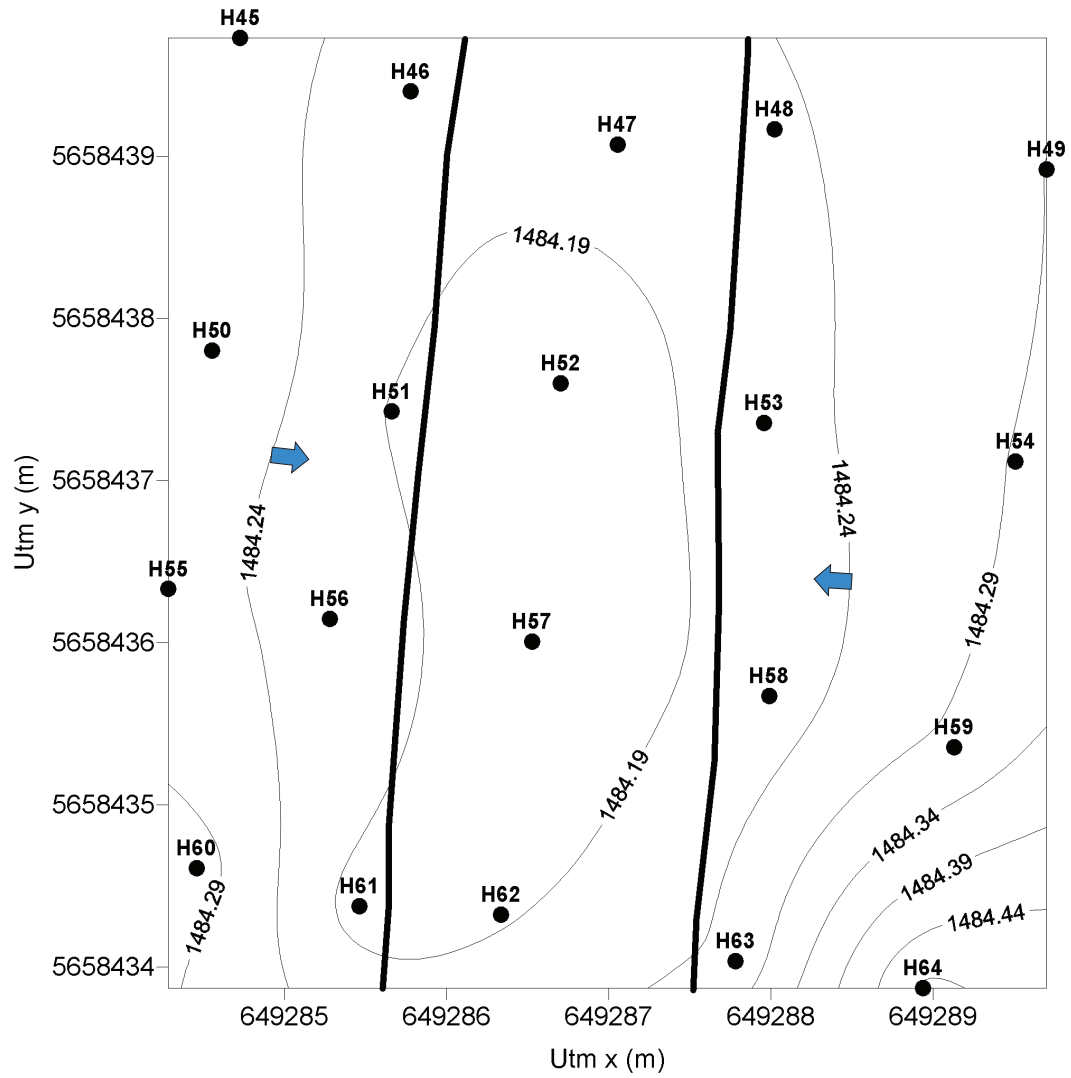
Flow nets for the reference reach show a similar pattern for 2006 and 2007: there was no indication of seasonal hyporheic flux in this portion of the study reach and on average, the stream is gaining water from the adjacent flood plain (Figures 5.12, 5.13, and C.3). This pattern of groundwater inflow to the stream was consistent throughout baseflow and stormflow conditions (data not shown).

The flow net (2006 seasonal mean) for the transverse transect above the north dam shows divergence of water away from the stream (Figure 5.14). The flux around the north dam ranges from  $\sim 0.002$  L/s on the west bank to  $\sim 0.009$  L/s on the east bank (Figure 5.20). Downstream of the dam we see a reversal of flow, where groundwater recharges the stream (Figure 5.15). Fluxes from the west bank here are small, but the east bank shows up to 0.02 L/s in 2006 (Figure 5.20).

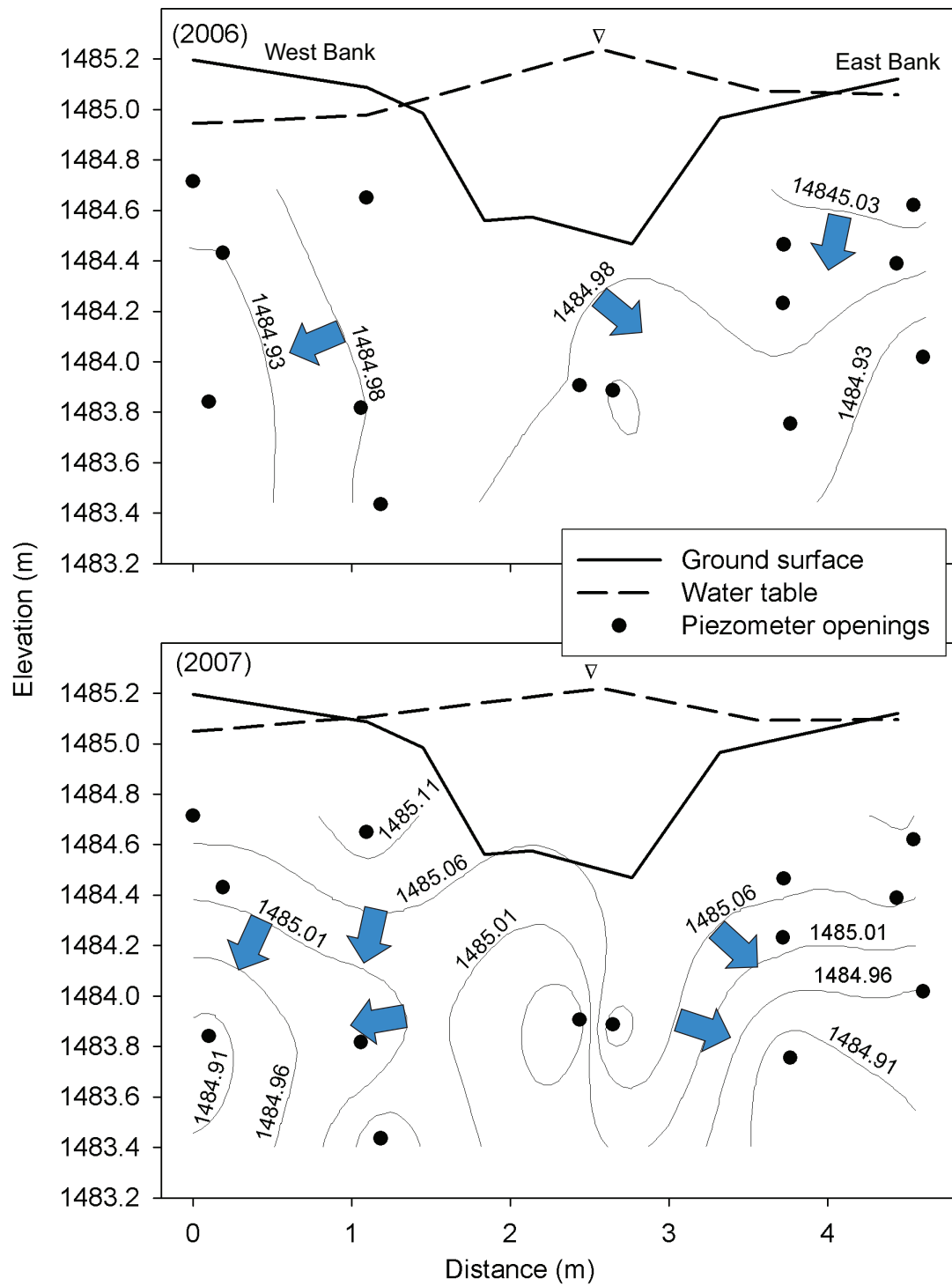
The symmetry in the flow pattern around the north dam in 2006 does not exist in 2007 (Figures 5.14 and 5.15). Above the north dam the water table mounds in the stream and drives flow horizontally from the stream to the aquifer in the nearby stream banks (Figure 5.14). Flux upstream of the dam in 2007 was  $\sim 0.001$  L/s on the west bank and 0.005 L/s on the east bank (Figure 5.20). Downstream of the dam, however, there is an asymmetrical flow pattern where the dominant flow to the stream is from the east bank (flux  $\sim 0.005$  L/s) (Figures 5.15 and 5.20). Flow on the west side of the stream has a less distinct pattern and fluxes are small (Figures 5.15 and 5.20); there is a convergence of flow in the piezometers nearest the west bank.



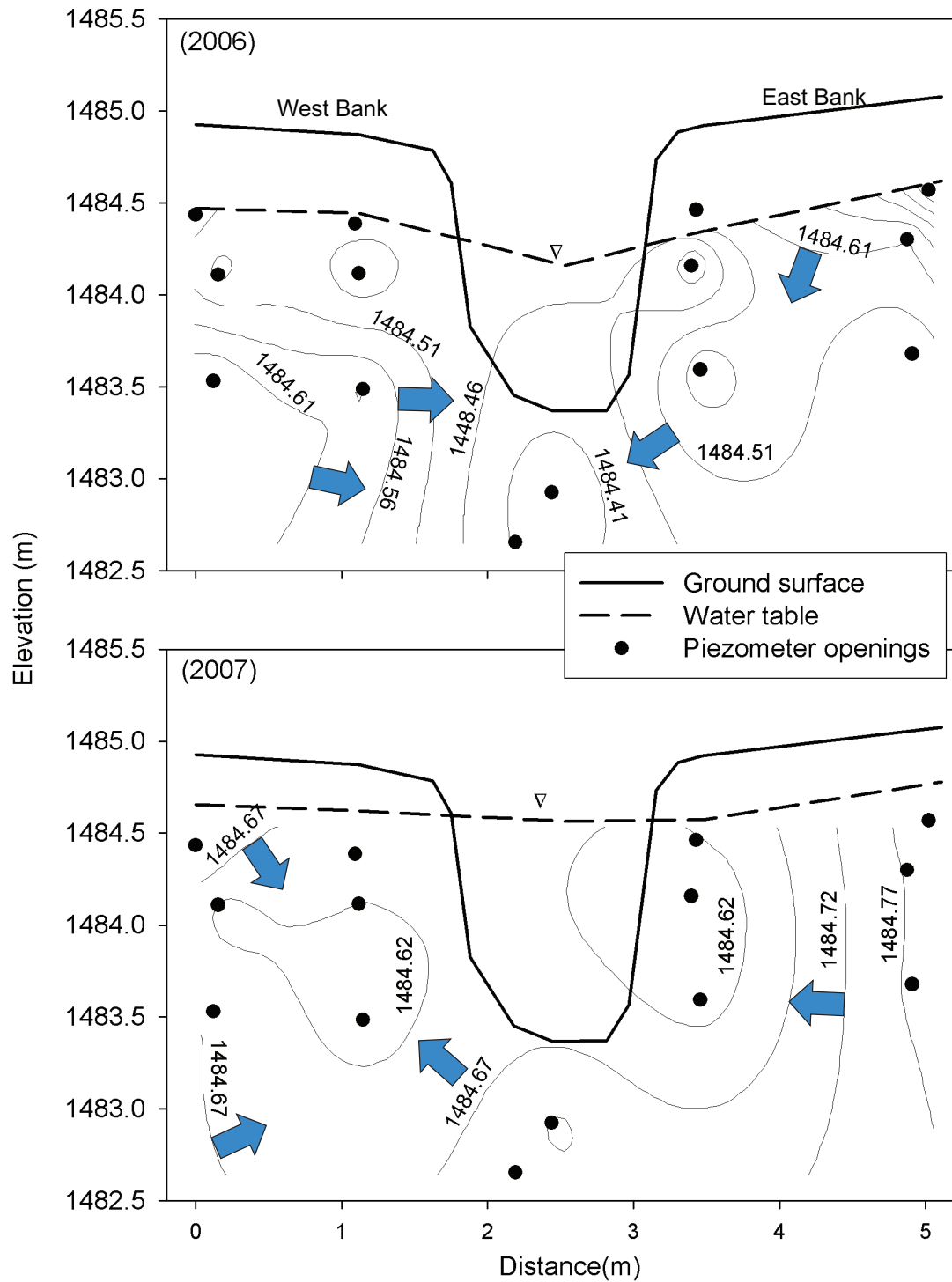
**Figure 5.12:** Reference reach plan view flow net of mean hydraulic head (m) for 2006



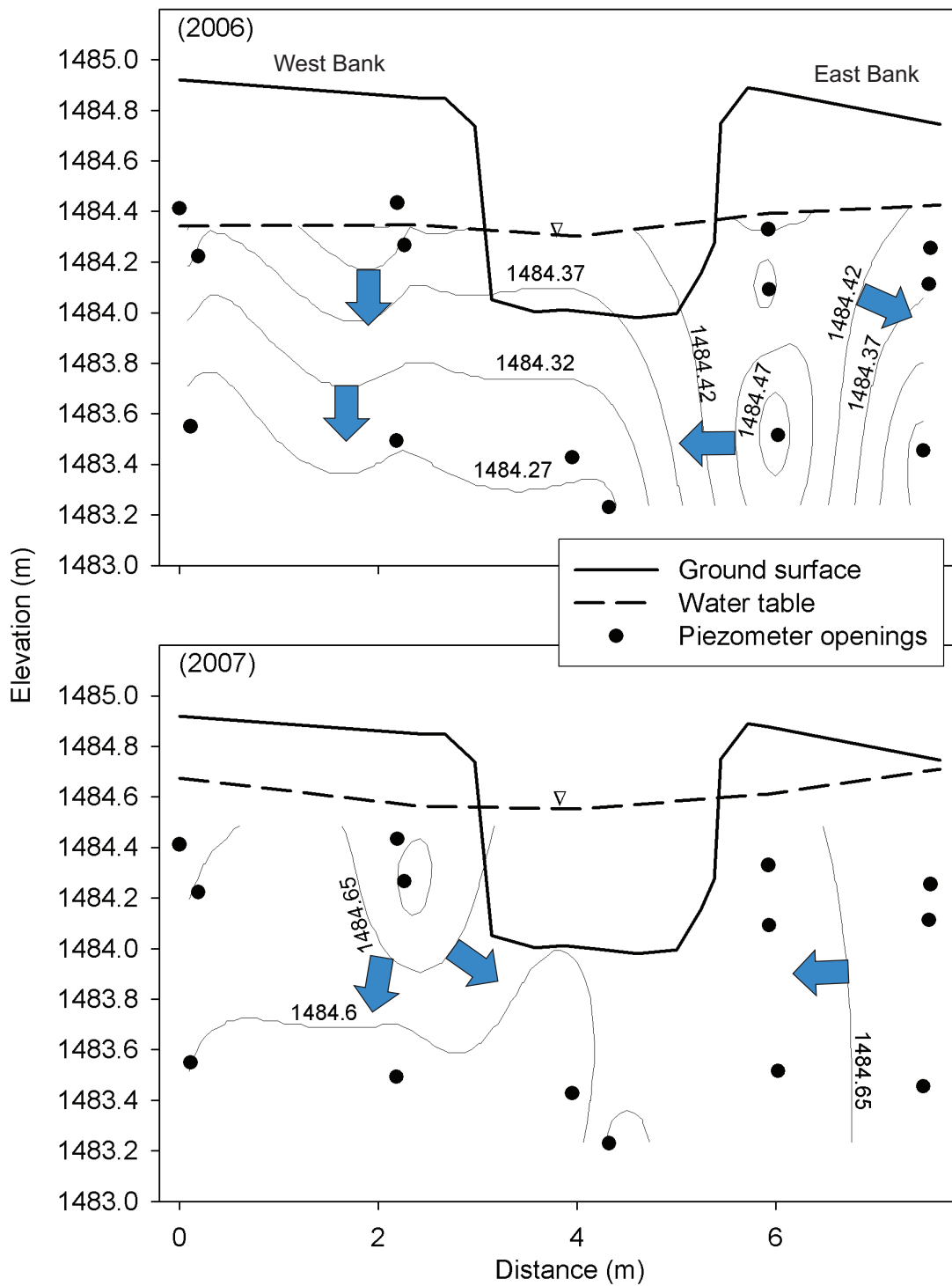
**Figure 5.13:** Reference reach plan view flow net of mean hydraulic head (m) for 2007



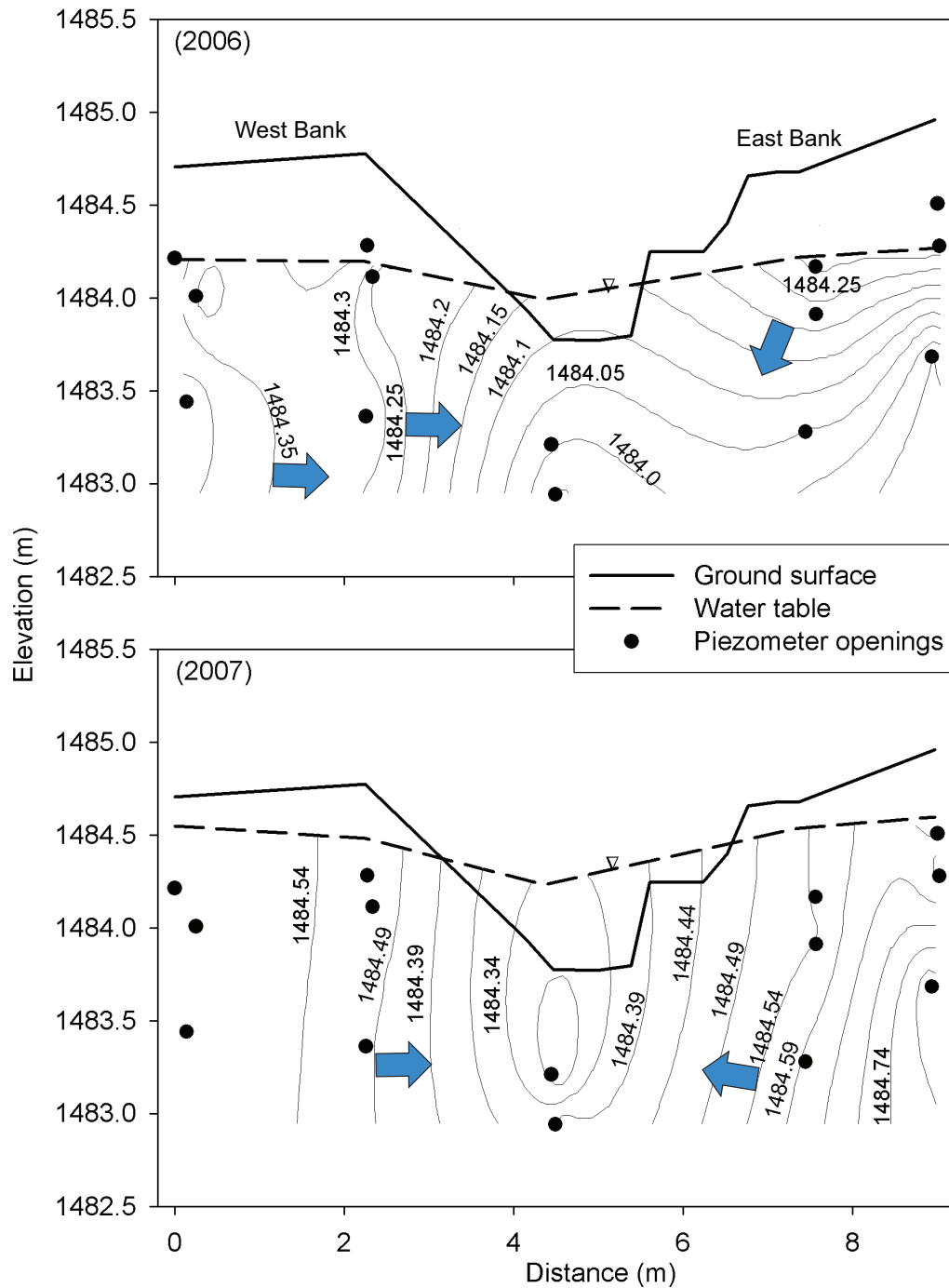
**Figure 5.14:** Mean piezometric head (m) upstream of the north dam for all eight sample dates in 2006 and 2007.



**Figure 5.15:** Mean piezometric head (m) downstream of the north dam for all eight sample dates in 2006 and 2007.



**Figure 5.16:** Mean piezometric head (m) upstream of the south dam for all eight sample dates in 2006 and 2007.

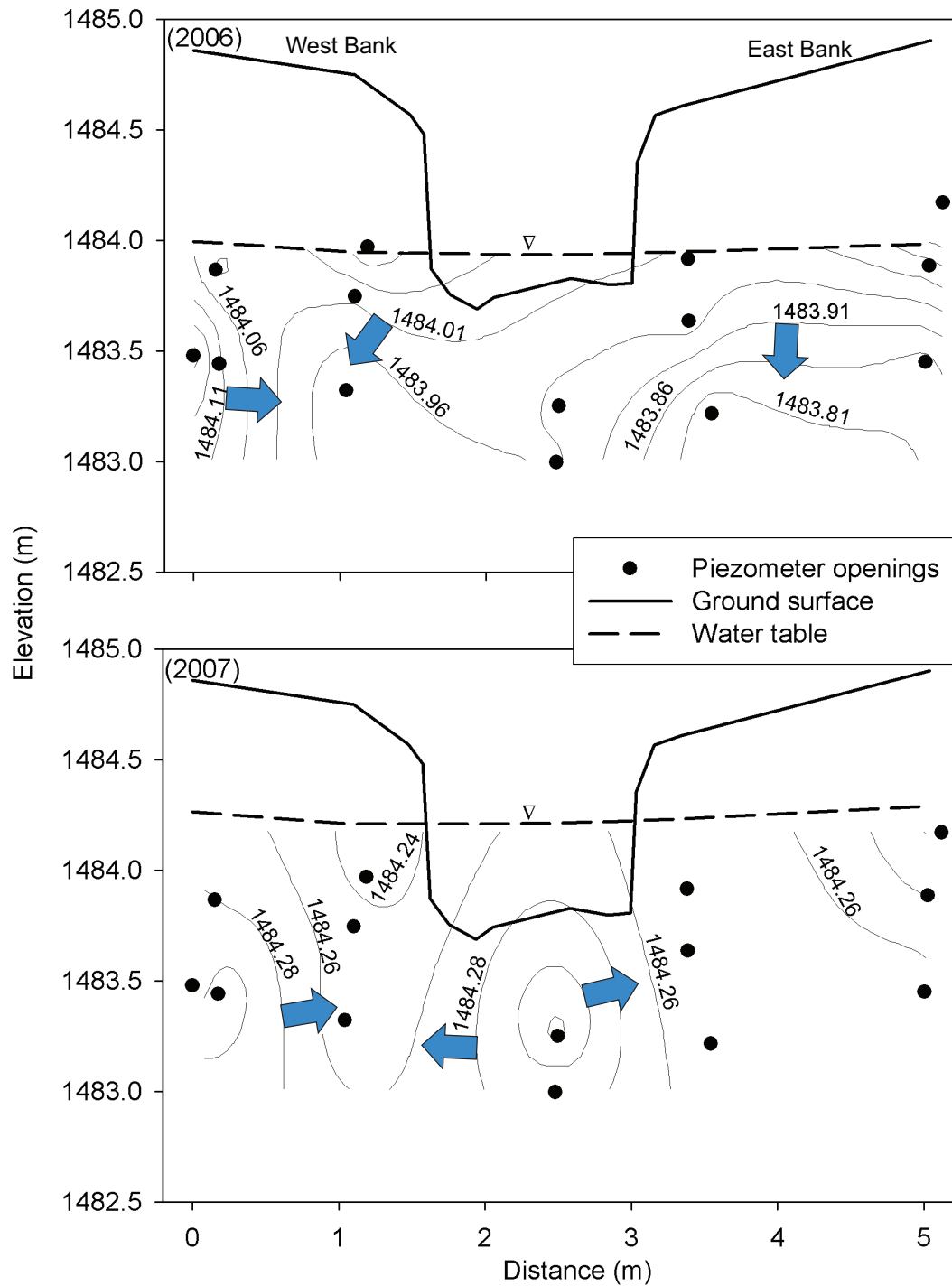


**Figure 5.17:** Mean piezometric head (m) downstream of the south dam for all eight sample dates in 2006 and 2007.

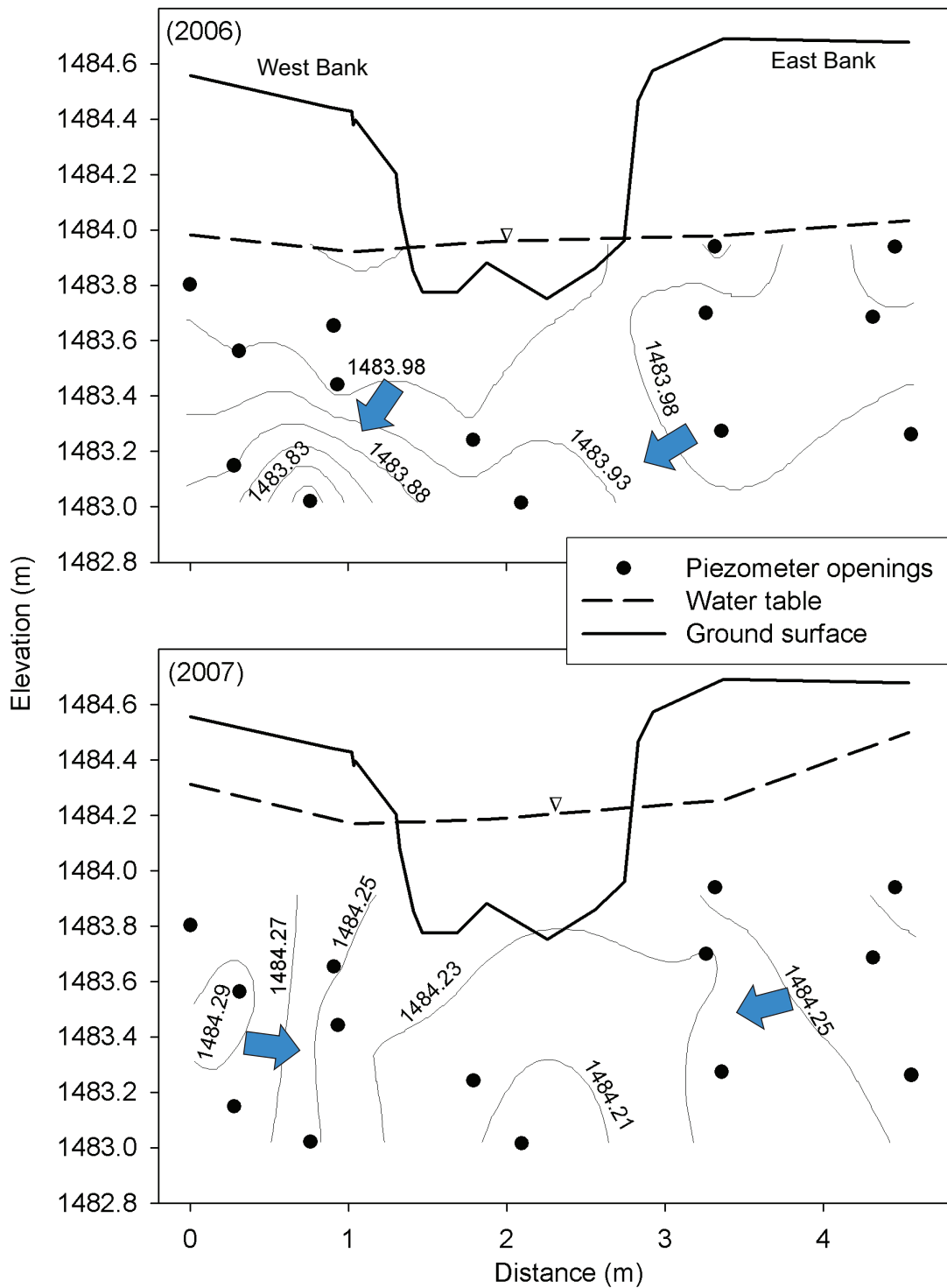


Unlike the conditions above the north dam, there was no clear evidence for hyporheic exchange upstream of the south dam in 2006 or 2007 (Figure 5.16). Hydraulic heads indicate that stream water is recharged from the east bank, but there is a downward flow of water in the west bank. Stream stage upstream of the dam was lower than the water table in both banks, indicating lateral flow towards the stream. Downstream of the south dam, the seasonal mean transverse flow nets indicate a potential for stream recharge in both 2006 and 2007, and fluxes that range from 0.001 L/s from the west bank to near zero from the east bank (Figures 5.17 and 5.20). The water table relative to the stream stage for both years also indicates that the aquifer recharges the stream (Figure 5.17).

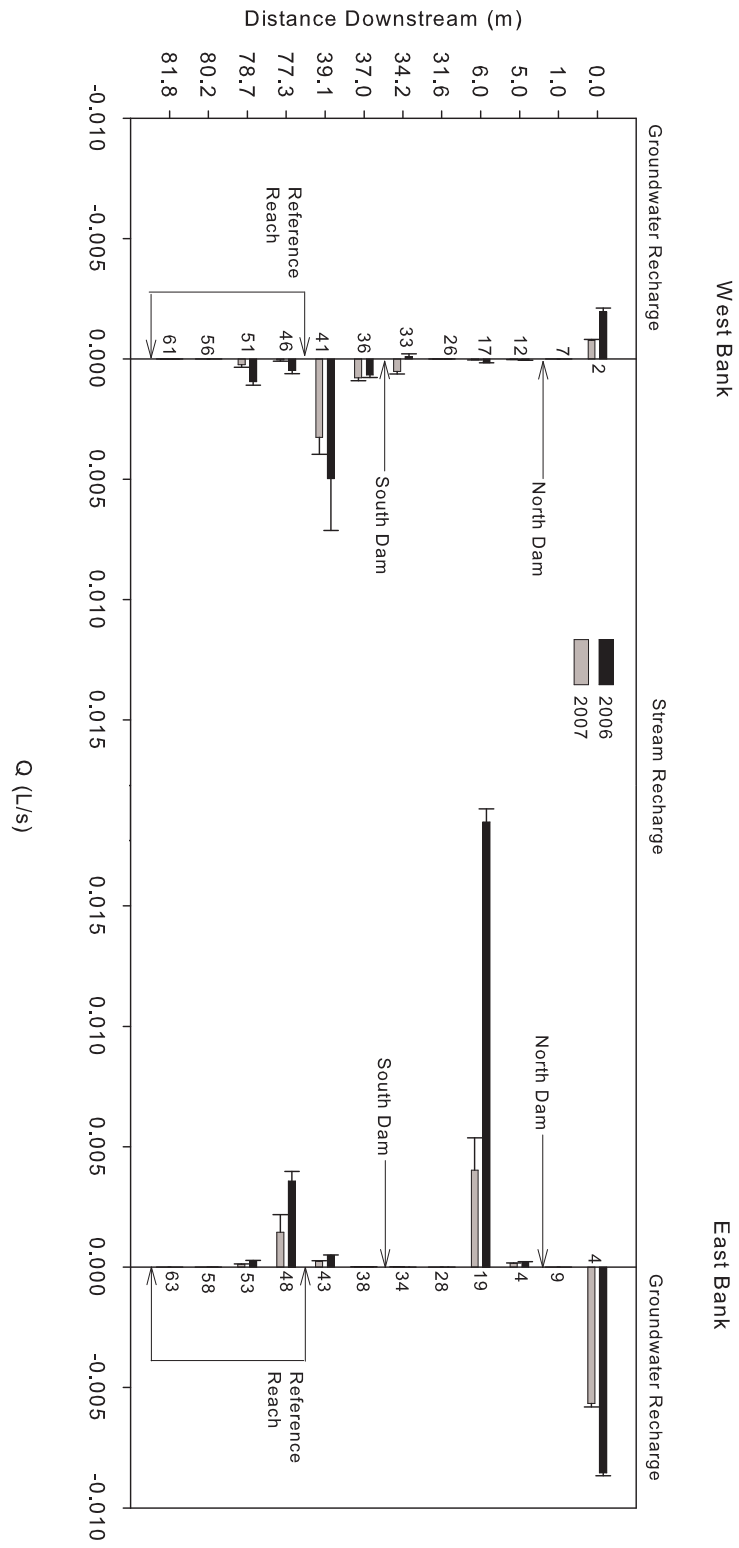
The transverse flow net for the reference reach in 2006 and 2007 shows the stream was gaining, with fluxes of  $\sim 0.001$  L/s and  $< 0.005$  L/s from the west and east banks, respectively (Figures 5.18, 5.19, and 5.20). The water tables for all four transects indicate static conditions, except in the most southern transect, in 2007 (Figure 5.19). Along this transect, the water table in the banks was higher than the stream stage, indicating a larger gradient for stream water recharge. The flow patterns for the reference reach do not indicate seasonal hyporheic exchanges.



**Figure 5.18:** Mean piezometric head (m) for the northern portion of the reference reach for all eight sample dates in 2006 and 2007.



**Figure 5.19:** Mean piezometric head (m) for the southern portion of the reference reach for all eight sample dates in 2006 and 2007.



**Figure 5.20:** Mean lateral hyporheic discharge ( $Q$ ) in 2006 and 2007.

# CHAPTER 6

## DISCUSSION

This examination of water exchange between a small beaver dammed stream and a shallow peat aquifer in the Canadian Rocky Mountains found that lateral hyporheic flows between the stream and adjacent peat aquifer were more dominant than vertical flows between the streambed and underlying aquifer. Instream beaver dams along the reach increased stream stage, creating seasonally stable VHGs and looped lateral transverse groundwater flow patterns around the north dam.

The first major finding of this study was that lateral hyporheic flow was more dominant than vertical hyporheic flow. Lateral hyporheic flows dominated in this system likely due to the difference in hydraulic conductivity between the bed and banks. The bank material was comprised mainly of peat, which had a hydraulic conductivity of  $K \sim 10^{-5} - 10^{-6}$  m/s, which was greater than the hydraulic conductivities of the instream piezometers ( $K \sim 10^{-8} - 10^{-9}$  m/s) (Figure 5.2), which represent the conductivity of the bed materials suggesting that the stream was underlain by fine silts, which impeded vertical flows. The control of hyporheic exchange by the hydraulic conductivities of the bed and bank material has also been observed in other studies (Triska et al., 1993b; Butturini et al., 2002; Wright et al., 2005). For example, Triska et al. (1993b) found that the permeability of the bankside sediments regulated the proportion of stream water and groundwater that was in the hyporheic zone in a small gravel-cobble stream in California. Greenwald et al. (2008) found that the vertical development of hyporheic flow in an arctic cobble stream with high hydraulic conductivity was greater than the vertical development in a nearby peat dominated stream with lower hydraulic conductivity, but similar size and characteristics. Further, modelling of hyporheic zones has shown hyporheic flow is highly

sensitivity to streambed and bank hydraulic conductivity (Wondzell and Swanson, 1996a; Lautz and Siegel, 2006). Other river systems where hydraulic conductivity of the bed and bank strata are more homogeneous may find the main channel flow gradient or valley floor width to be the limitation on flux (Wondzell and Swanson, 1996a).

The dominance of lateral flows over vertical flows in this system differs from the findings of Wondzell and Swanson (1996b), who worked in a stream with pool-riffle-pool sequences and found that lateral fluxes were negligible either due to failure in the well networks ability to observe such flow paths or the steep potential gradients found between the stream and a secondary channel. Lautz and Siegel (2006) also found vertical flow dominating over lateral flow in a arid mountain stream due to the hydraulic gradients and the large changes in stream stage associated with dams. In another study Lautz et al. (2006) found that lateral exchange did take place, but level to low groundwater tables next to higher stream stage created flow paths that sloped laterally away from the stream and did not return to the stream in short transient flow paths, but were lost to longer hyporheic time-scales.

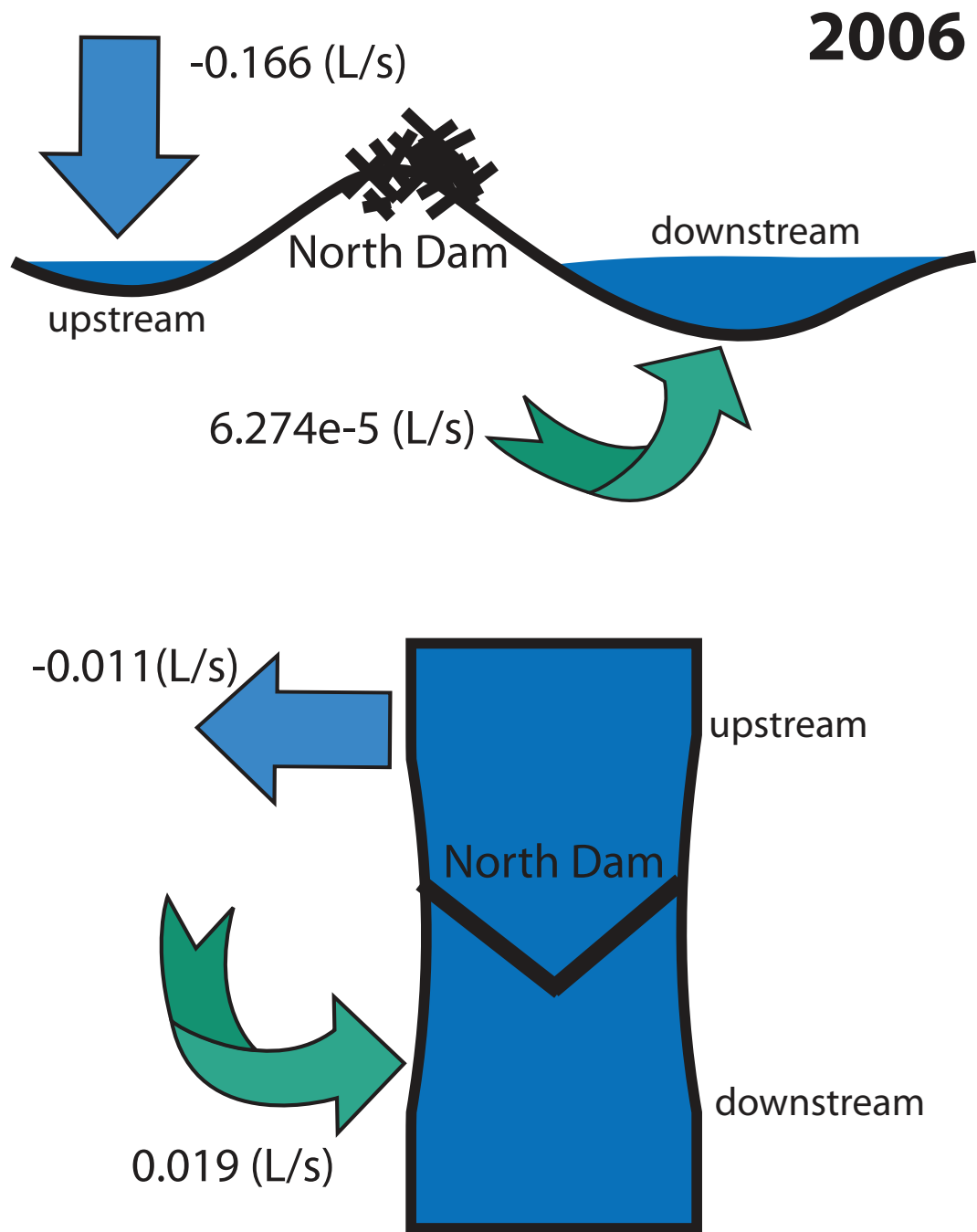
The second important finding in this study is that little hyporheic exchange (vertical or lateral) was found along the reference reach where there was an absence of beaver dams. In contrast portions of the stream with instream beaver dams had enhanced vertical and lateral hyporheic fluxes as well as altered groundwater flow paths in the near streambed/bank environment.

The hydraulic gradients in the reference reach did not indicate flow reversals that would be expected during high and low stream stage events. Bank storage theory (Figure 2.1) states that during high stream stage events flood waters should propagate in to the banks and recharge the bank aquifer. This stored water should then discharge back to the stream during the streamflow recession. Results from the reference reach, however, do not support this conceptual model. Bank recharge during high stream stage events and reversed bank discharge during low flow was absent. Instead, flow was directed toward the stream in the reference reach during the entire study. For instance, the hydraulic gradient between the near bank and

stream on 13 June 2007, the day of the highest stream discharge sampled (Figure 5.1), was directed toward the stream throughout the reference reach and  $\Delta h$  for the west bank ranged from 0.013 to 0.097 m and  $\Delta h$  for the east bank ranged from 0.041 to 0.104 m. The lowest discharge occurred on the 26 June 2007 (Figure 5.1) and head gradients indicated flow was still directed toward the stream, except at the northern most transect. Gradients on this day were very similar to those measured during the high discharge event on 13 June 2007:  $\Delta h$  for the west bank ranged from -0.064 to 0.097 m and  $\Delta h$  for the east bank ranged from -0.053 to 0.281 m. These results show that during both high and low flow events groundwater discharge from the bank to the stream is prevalent in this portion of the study reach.

Around the north and south beaver dams, VHGs in the shallow A piezometers showed a distinct pattern of flow looping that was not observed in the undammed portions of the study reach. This pattern of stream water movement into the bed upstream of the dams and out of the bed downstream of the dams persisted during both low or high flows. Strong vertical gradients were created by the impoundment of water upstream of the dam, which created greater hydraulic heads forcing water down into the streambed. Downstream of the dam the lower stream stage created an area of low hydraulic head and the potential for water to come back into the stream. While other researchers have also found similar vertical looping flow patterns beneath dams (White, 1990), very little water actually returned to the stream downstream of the north or south dam (Figures 6.1 and 6.2). Instead, the data suggest stream water vertically recharges the underlying aquifer upstream of the dams and then is either lost to the groundwater flow system or returns back to the stream in the reference reach (Figure 6.3 (2007)).

The transverse transect flow nets illustrated how the north dam created lateral hyporheic flow pathways and altered the groundwater flow patterns in the adjacent bank(s). The beaver dams elevated stream stages upstream of the dam, moving water out to the adjacent aquifer where the water table was lower than the stream stage. The flow gradient was then reversed downstream of the dam where groundwater mounds were found to recharge the stream. The pattern of hyporheic exchange

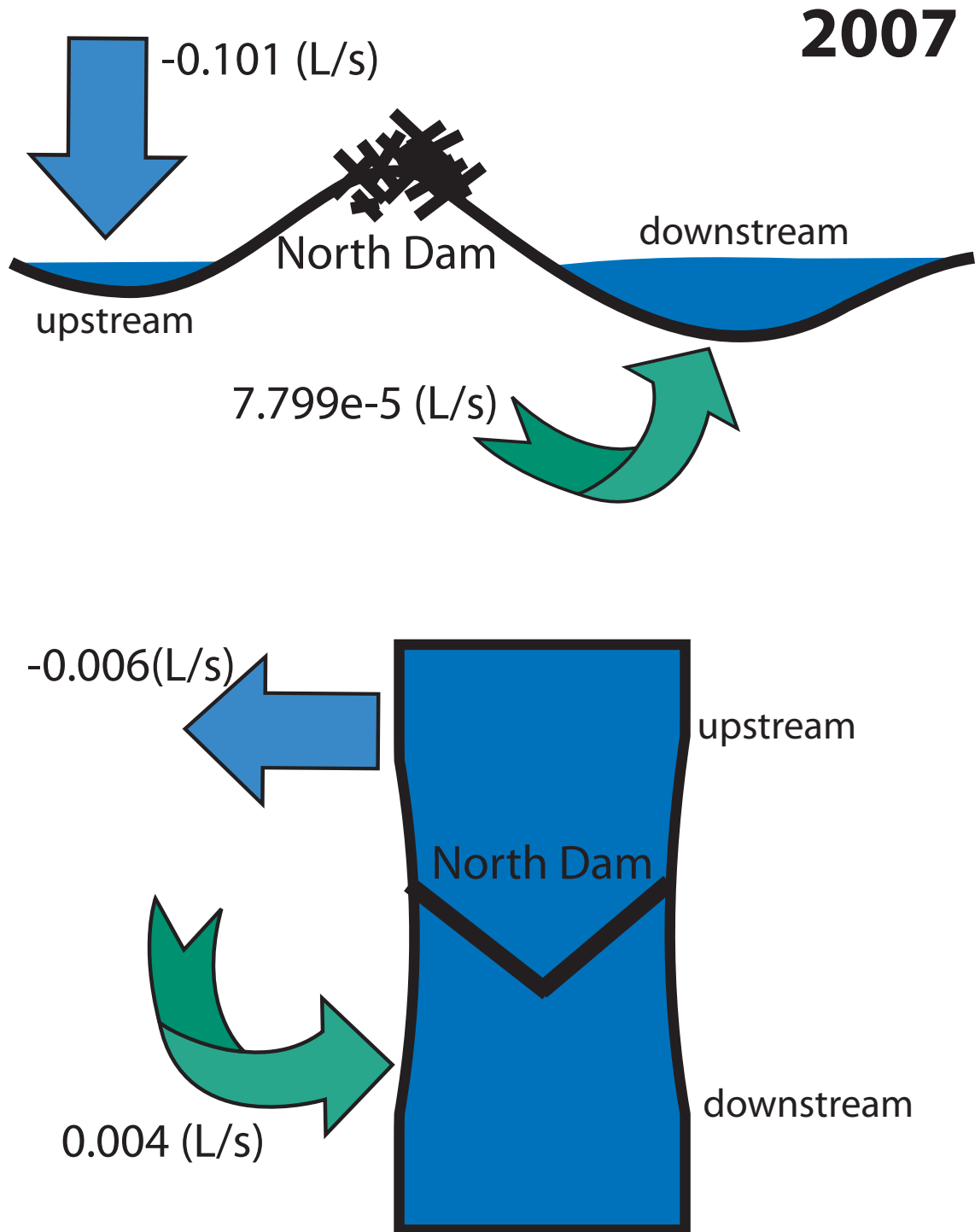


**Figure 6.1:** Mean comparisons of vertical hyporheic flux to lateral hyporheic flux (from both banks) directly upstream and downstream of the north dam in 2006.

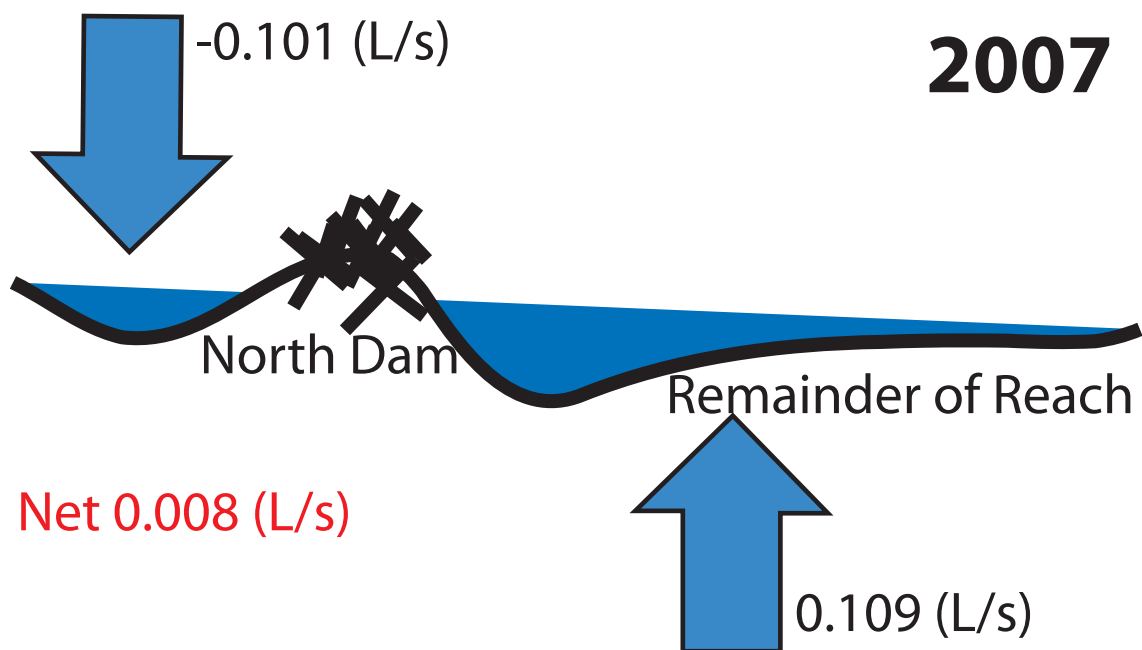
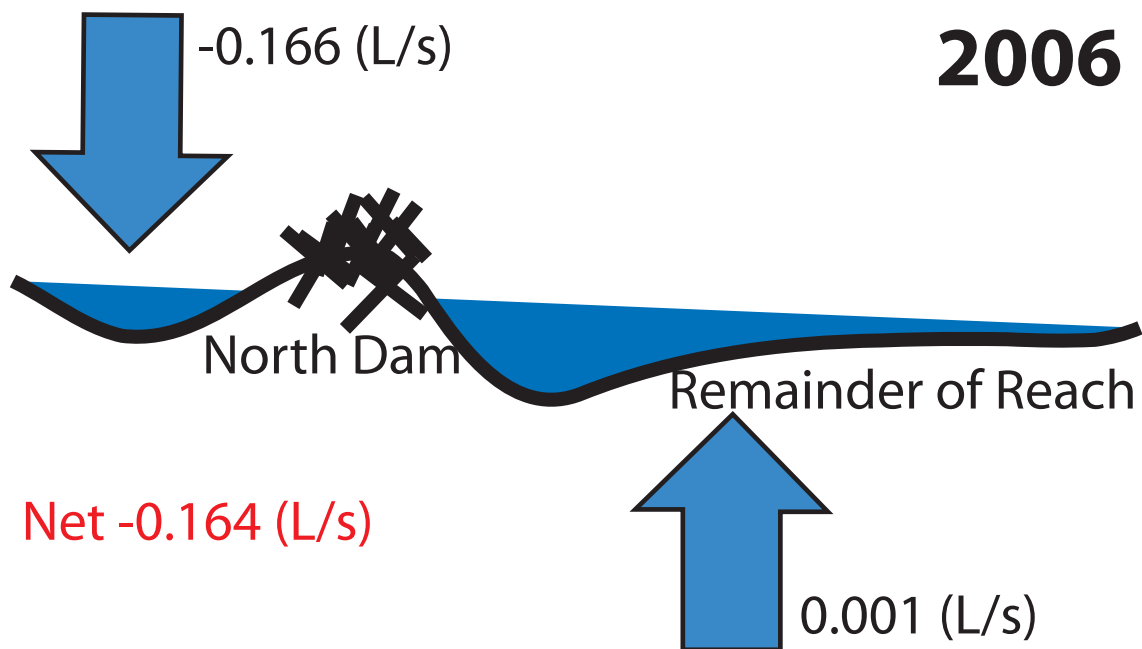


around the north dam was not observed around the south dam or in the undammed reference reach. It is likely that the head difference in the stream near the south dam was insufficient to generate hyporheic exchange. The gradient between the stream and bank aquifer near the south dam was small. This suggests a threshold dam size exists in order to generate looping hyporheic flow paths around dams in this environment. For this study, the dam height threshold was between 0.67 m and 1.05 m. Other studies have found that, within gaining reaches of an arid mountain stream, dams as low as 0.5 m were sufficient to generate vertical hyporheic exchange (Lautz and Siegel, 2006). Hill et al. (1998) researched lateral hyporheic exchanges between pool riffle sequences in a third order stream in southern Ontario and found very low slope thresholds 2 – 3% for riffles and surface water slopes as low as  $< 1\%$  for pools were needed. Future studies should further explore this idea of a threshold beaver dam height to generate hyporheic exchange and how bed and bank substrate type affect it.

Observations of VHG for every day sampled illustrated that instream dams created seasonal stability in the hydraulic gradients so that hyporheic fluxes are close to constant over time, regardless of stream stage. Wondzell and Swanson (1996a) also found that overall patterns of hyporheic flow changed little over the course of a year in a fourth order mountain stream and attributed their findings to the longitudinal gradient of the main channel (decreased exchange with increasing stream order) and the influence of secondary channels rather than dams. At this site, the instream beaver dams created stable hydrological head conditions in the stream and bank environment, which kept hydraulic gradients and hyporheic flows relatively constant. In areas that were undammed, however, the bank environment was more prone to drying out as seen by a lowering of the water table over the summer, which restricted surface and groundwater exchanges. These data thus provide support for Westbrook et al. (2006) finding that beaver dams are key in maintaining wetland conditions in the riparian zone throughout both high and low streamflow conditions.



**Figure 6.2:** Mean comparisons of vertical hyporheic flux to lateral hyporheic flux (from both banks) directly upstream and downstream of the north dam in 2007.



**Figure 6.3:** Mean vertical stream discharge loss from upstream of the north dam versus the total mean stream recharge gain throughout the remainder of the reach for 2006 and 2007.

# CHAPTER 7

## CONCLUSIONS

A wide peat dominated valley in the Front Ranges of the Canadian Rocky Mountains did show that small (0.67 to 1.05 m high) instream beaver dams do enhance vertical and lateral hyporheic exchange and alter groundwater flow patterns in the near stream sediment environment. The results from this study determined that lateral hyporheic fluxes dominated over vertical fluxes on a 90 m reach of a small beaver dammed stream. Lateral flow paths around the north dam took place within 6 m directly upstream and downstream of the dam in the near bank sediments. The hydraulic conductivities of the stream bank (peat) versus stream bed (silt) material was the controlling factor in the development of transient hyporheic flow paths. Vertical fluxes from the stream to groundwater upstream of the largest dam suggested that this water was lost to groundwater recharge that may have returned to the stream further down the stream reach. Beaver dams  $> 1.05$  m were found to create and maintain seasonally stable looping flow gradients and this pattern was not observed in the undammed portion of the reach. The undammed reference reach showed gaining stream conditions at high and low stream stages suggesting that any groundwater recharge observed in other parts of the stream were caused by the beaver dams.

The hyporheic fluxes measured laterally in the study reach demonstrate, on a small scale, the potential for beaver to create seasonally stable hydrological connections that link the stream to the bank and riparian zone. Within the Sibbald Research Basin there are numerous large dams (20 – 100 m or longer and 0.5-2.5 m high) that have very large ponds associated with them. The implication of this work for valley scale flow patterns is that the larger beaver dams have the potential to con-

nect the whole valley bottom to the stream. These findings can be further extended to valleys within the Canadian Rocky Mountains where beaver are prevalent due to ample supplies of willow. Many of these valley bottoms may be wetlands simply because beaver may be maintaining riparian water tables near or at the ground surface by creating large-scale (valley sized) hyporheic flow pathways. Such hydraulic connectivity may persist throughout wet and dry periods as beaver keep hydrological conditions relatively static during the summer period (Westbrook et al., 2006).

During flood conditions, theories such as bank storage and flood pulse concepts will likely not apply because in these systems the banks are already saturated meaning storage is already at capacity. Thus, we would expect to see quick response times to precipitation events. A ramification would be that flood models would have to be adapted to accurately predict flood responses from beaver dammed river systems.

In addition to the impacts beaver have on hydrological flow patterns, water quality and biogeochemical reactions are also important processes affected by hyporheic exchange (Triska et al., 1993b). Hyporheic exchange zones have been noted as sources for nutrients in nutrient limited streams improving overall stream ecology (Wondzell and Swanson, 1996b). Alternatively Hill and Lymburner (1998) found that in agricultural areas where streams are nutrient rich hyporheic zones can be nutrient sinks and used in stream restoration projects. Beaver dam enhanced hyporheic flow then has the potential to regulate biogeochemical transformations and factors influencing water temperatures, which then help maintain healthy ecosystems within riverine systems (White, 1990).

In the pursuit of maintaining and managing our water resources, understanding the physical, chemical and biological processes taking place within our aquatic environments becomes essential. Combining and studying these processes together helps to create a holistic conceptual model of the way in which water travels through valley systems and improves our overall understanding.

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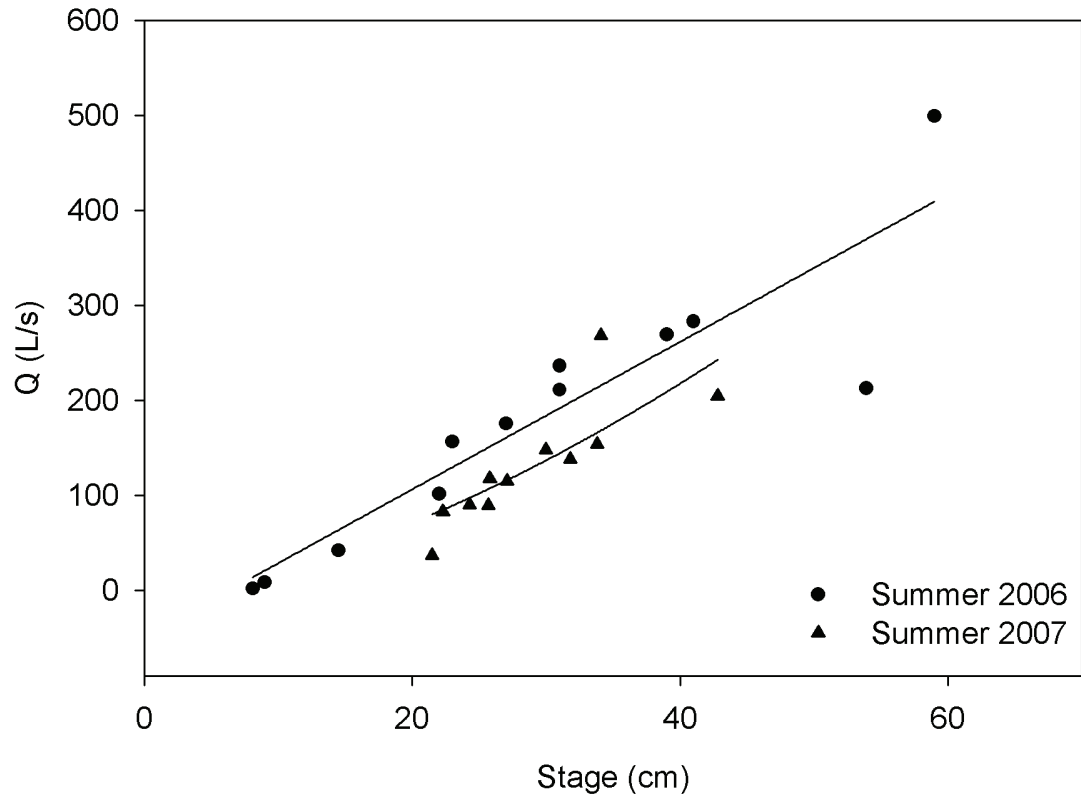
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# APPENDIX A

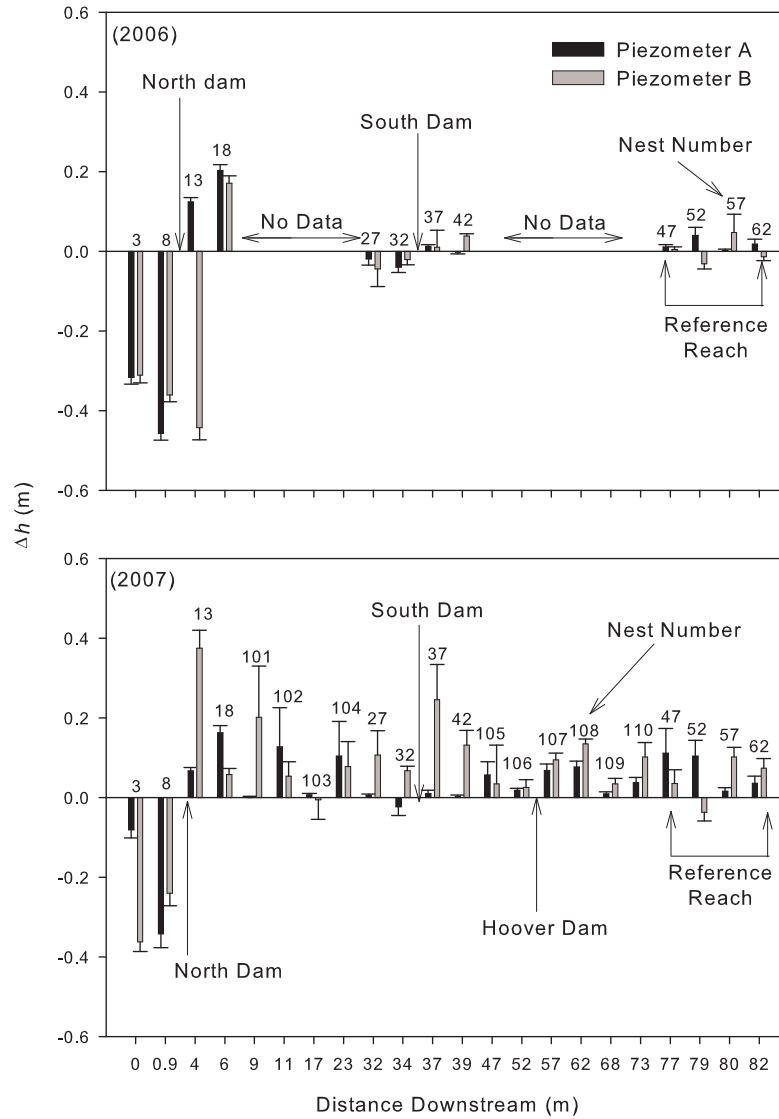
## RATING CURVES



**Figure A.1:** Stream discharge rating curves for 2006 with an  $R^2$  value of 0.82 ( $y = 7.76x - 48.63$ ) and 2007 with an  $R^2$  value of 0.74 ( $y = 0.83x^{2.17}$ ).

# APPENDIX B

## VERTICAL HYDRAULIC HEADS



**Figure B.1:** Seasonal mean instream  $\Delta h$  for 2006, 2007 piezometers A and B.

# APPENDIX C

## TOPOGRAPHIC CONTOUR PLOTS

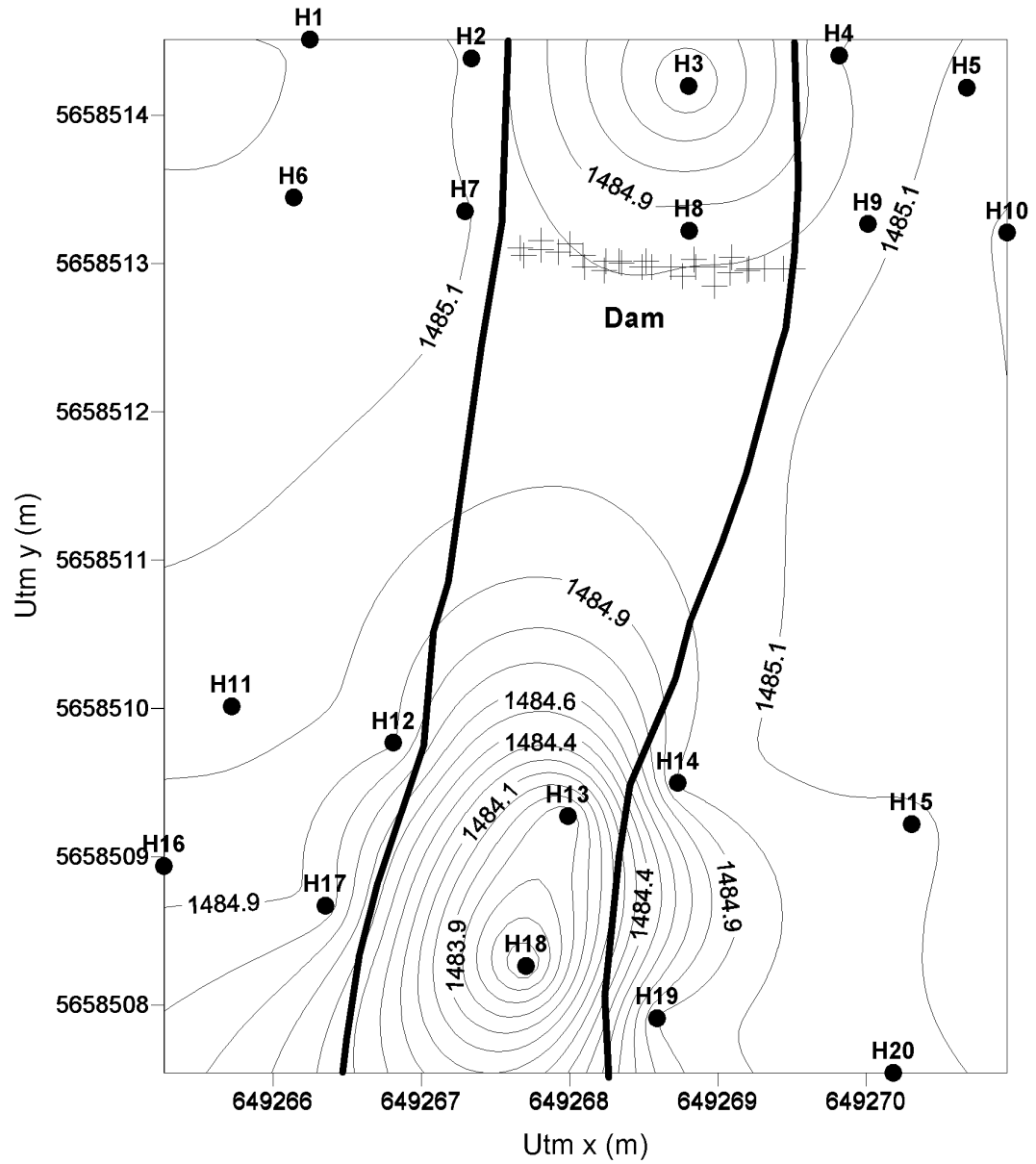


Figure C.1: North dam plan view of contoured ground elevation

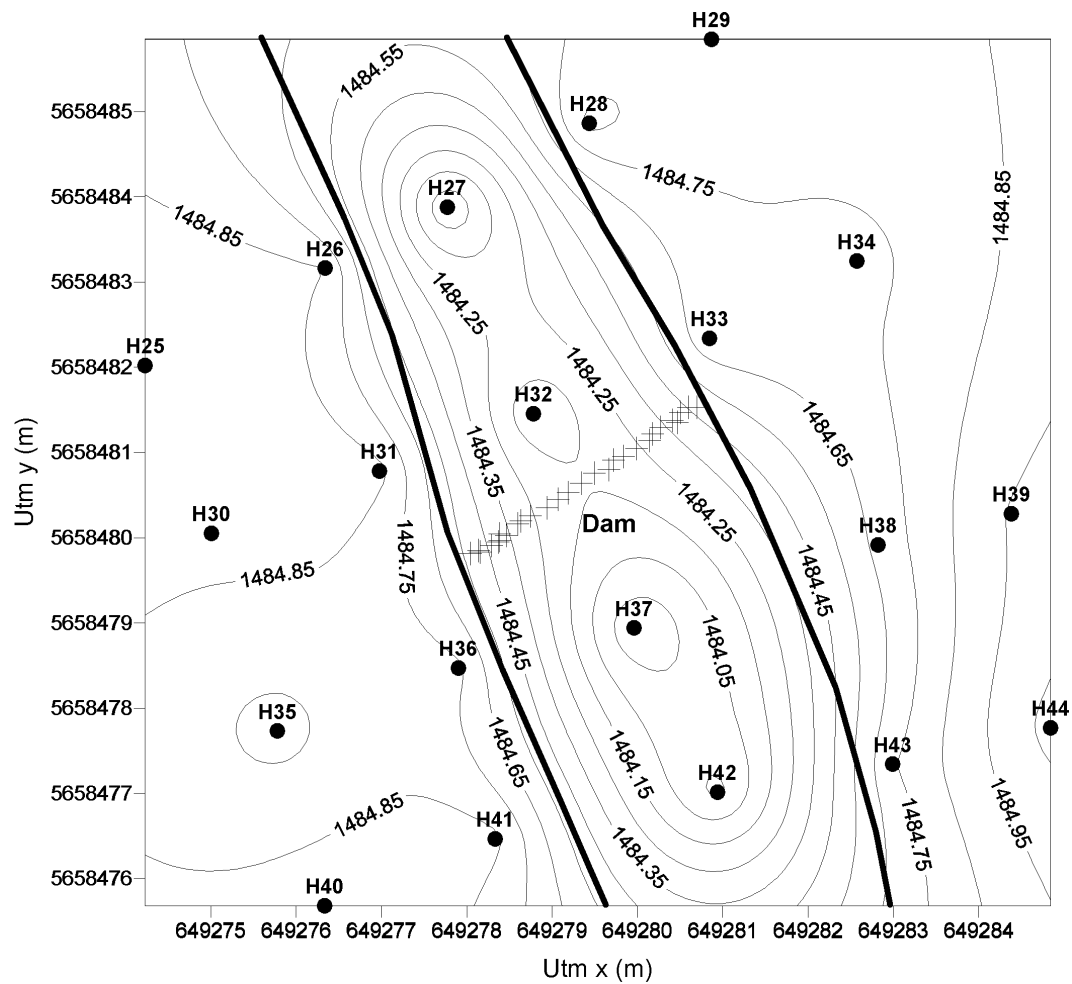


Figure C.2: North dam plan view of contoured ground elevation

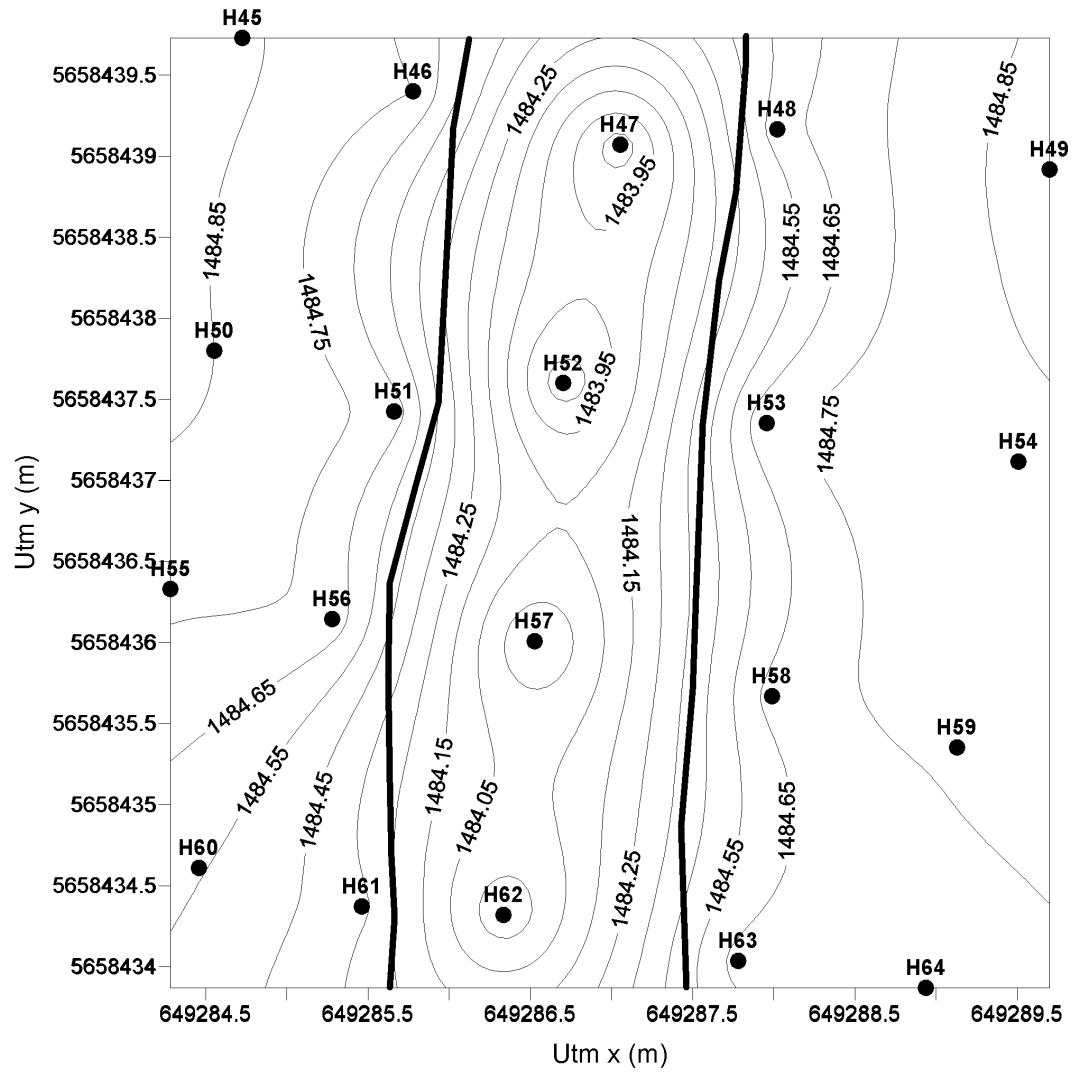


Figure C.3: Reference plan view of contoured ground elevation